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HUNGARIAN ACADEMY OF SCIENCES

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PREFACE

Organized by the Geographical Research Institute of the Hungarian Academy of Sciences and the Hungarian National Committee of the International Union for Quaternary Research, a paleogeographical seminar was held in Tata, Komárom county, between 16-22, October, 1986. This was an important event in the cooperation of long tradition between the Polish and Hungarian Academies of Sciences. A Polish delegation of five members, led by Professor Leszek Starkel, arrived to Hungary to discuss with Hungarian colleagues the problems of reconstructin Upper Pleistocene and Holocene environments on both sides of the Carpathians. The central issues were approached from paleontological, archaeological, landscape ecological, geomorphological, sedimentological and geochronological aspects. The present proceedings volume includes the Polish contributions and papers based on the field representations by Hungarian experts in the Transdanubian mountains and lowlands in the topics of loess and terrace research, various chronological aspects and earlier and present-day geomorphic processes. The volume is meant to illustrate the interdisciplinary approach to the reconstruction of Quarternary environments.

Budapest, July 1988

Márton Pécsi
ordinary member
Hungarian Academy of Sciences

QUATERNARY EVOLUTION IN HUNGARY

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NEOGENE AND QUATERNARY GEOMORPHOLOGICAL SURFACES AND LITHOSTRATIGRAPHICAL UNITS IN THE TRANSDANUBIAN MOUNTAINS

M. Pécsi, Gy. Scheuer, F. Schweitzer

ABSTRACT

In the Hungarian Mountains, particularly on the Transdanubian side of the Danube valley, the geomorphological surfaces preserved by travertines (6-7 fluvial terraces and the higher-lying one or two pediments and two or three raised beaches) are unique records of events in geomorphic evolution. The reconstruction of denudation chronology for the Quaternary and late Tertiary was based on the evidence supplied by these surfaces.

INTRODUCTION

Along the Hungarian section of the Danube valley, in limestone and dolomite mountains affected by the Alpine orogeny, 6-7 fluvial terraces, at higher levels additional 4-5 geomorphological surfaces (pediments and raised beaches) occur in a step-like fashion. Most of these geomorphological surfaces have been prserved by travertine caps.

In the early 1970's a working group under the guidance of M. PÉCSI was set up to suggest a denudation chronology of the travertine-capped levels. Partial achievements of this research have already been published, emphasizing the conditions of travertine genesis, as in the collection of papers issued for the 12th INQUA Congress (PÉCSI, SCHEUER, SCHWEITZER in: PÉCSI, 1982). The results of geochronological investigations have also appeared at regular intervals (KRETZOI, PÉCSI, 1979, 1982; PÉCSI, 1973, 1975; SCHEUER, SCHWEITZER, 1970, 1973, 1974, 1979, 1980, 1983; PÉCSI, SCHEUER, SCHWEITZER, 1978, 1982, 1984; PÉCSI et al. 1985).

During the years terrace and travertine datings have accumulated and, consequently, in addition to the geological-geomorphological position, more and more biostratigraphical (faunal) data, paleomagnetic and radiometric (C^{14}, Th/U and ESR) information have become available for evaluation (HENNIG et al., 1983; PÉCSI, OSMOND, 1973; SCHEUER,SCHWEITZER, 1974; JÁNOSSY, 1979; KRETZOI, VÉRTES, 1965; PÉCSI et al. 1985).

One of our main targets was to reconstruct a denudation chronology for the Neogene and, particularly, for the Quaternary evolution and geomorphological surfaces and to identify lithostratigraphical units in key sections.

It is very characteristic of the Hungarian Mountains that the presence of Neogene and Quaternary geomorphological surfaces is emphasized by their travertine caps, preserving them (especially the older ones) from denudation. Even after some simplification, in the Danube Bend mountains at least 12, lithostratigraphically distinct, travertine horizons are observed.

The frequent occurrence of travertine on terraces had encouraged researchers to claim that travertines mostly deposited from karst springs issuing at the base level of erosion (SCHRÉTER 1953; PÉCSI 1959).

Recently, our investigations have supported this view in many respects; however, some travertine formations deposited in valley sides or on terraces, instead of flood-plains, arranged in a step-like fashion (tetarata forms - SCHEUER, SCHWEITZER, 1973, 1979). This indicates that these tetarata type travertines may have developed on major terraces almost simultaneously. We observed this phenomena primarily on some higher terraces, accompanied by the occurrence of thermal spring vents and cones (SCHEUER, SCHWEITZER, 1973, 1983).

Also by detailed lithostratigraphical analyses the long-term, multiphasal development of some travertine series has been proved (SCHEUER, SCHWEITZER, 1983). Thus, not all travertine series have formed in a single period like an interglacial or interstadial. In thicker series sand or sandy loess intercalations, fluvial sand layers, paleosols,alluvial debris or deluvial deposits are equally frequent (in some 20-40 m thick travertine series 3-4 loess layers may be found). This allows the conclusion that travertine formation was interrupted for shorter of longer intervals (SCHEUER, SCHWEITZER, 1974).

For easy reference to the various surfaces, symbols are used to identify them: tI - tVII for terraces and T1-T7 for the travertines overlying them. For the higher levels the following symbols are used: PI-PII for pediments, mI-mIII for raised beaches. The travertines capping the latter are designated as T8-T12. The lowest number always refers to the youngest formation, while the highest for the oldest. This is only a relative chronology and only valid for an individual section, in other cases only helps orintation and comparison.

We intended to find typical representatives for these levels and describe them as type localities, where lithostratigraphical units are distinct.

The time-scale applied is in accordance with international proposals (Plio-Pleistocene boundary is at 1.8 ma BP, Mio-Pliocene boundary at 5.4 ma BP).

QUATERNARY TERRACES AND TRAVERTINES

The youngest (late Holocene) travertine formation (T1) deposited on the flood-plain of the Tata river from karst springs (Fig. 1). After its locality it is called Fényes-forrás Travertine.

For the absolute dating of the higher floodplain level of the Danube, the nearby 'Almásfüzitő Danubian gravel' was used. The radiocarbon age of a carbonized log found 6 m deep in the flood-plain deposit is 11,850±110 a BP (HV 6958), i.e. postglacial (Fig. 2).

The Római-fürdő Travertine Formation (T2) overlying the first flood-free terrace (no II/a) has a geomorphological-geological position pointing to early Holocene, postglacial age (Fig. 3). This is 1-2 m thick, of loose structure and restricted to the environs of the former spring lake. its base is the sandy gravel of the Danube (at 105 m above sea level).

The second flood-free terrace (no II/b) of the Danube is covered by the Tata Travertine (T2b). This travertine formation (Fig. 4) encloses the Moustérian Tata culture (VÉRTES et al. 1964). The finds in this 4-6 m thick travertine series were recovered from a sand lens filling a

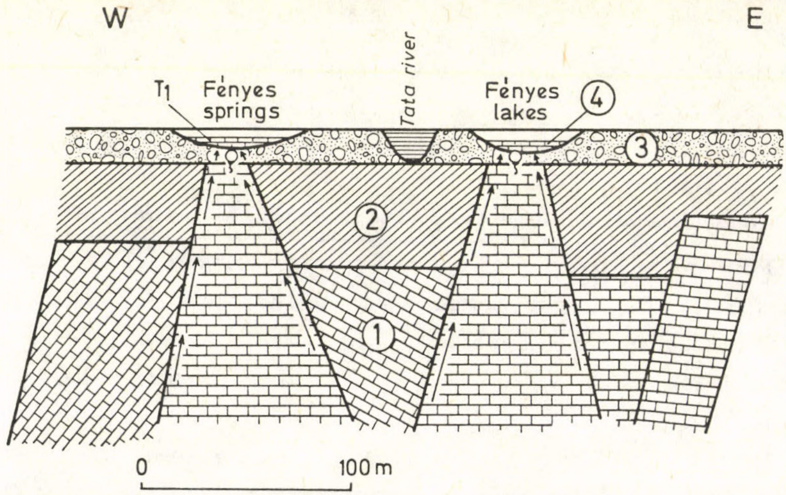


Fig. 1: The Fényes-forrás Travertine (T1) Holocene flood-plain travertine formation associated with springs issuing from covered horst through fluvial deposits (after SCHEUER, SCHWEITZER).
1 = Triassic limestone; 2 = Tertiary sediments; 3 = recent deposits of the Tata river; 4 = lacustrine-marshy type of travertine

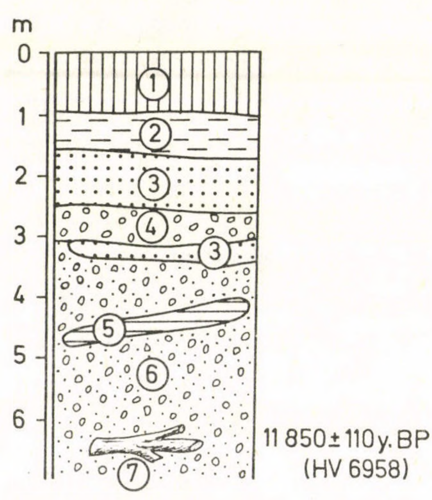


Fig. 2: The Almásfüzitő Gravel, postglacial, higher flood-plain level gravel of the Danube (tI - after PÉCSI, SCHWEITZER). 1 = chernozem; 2 = silt; 3 = fluvial sand; 4 = gravel; 5 = clay; 6 = sandy gravel; 7 = carbonized log

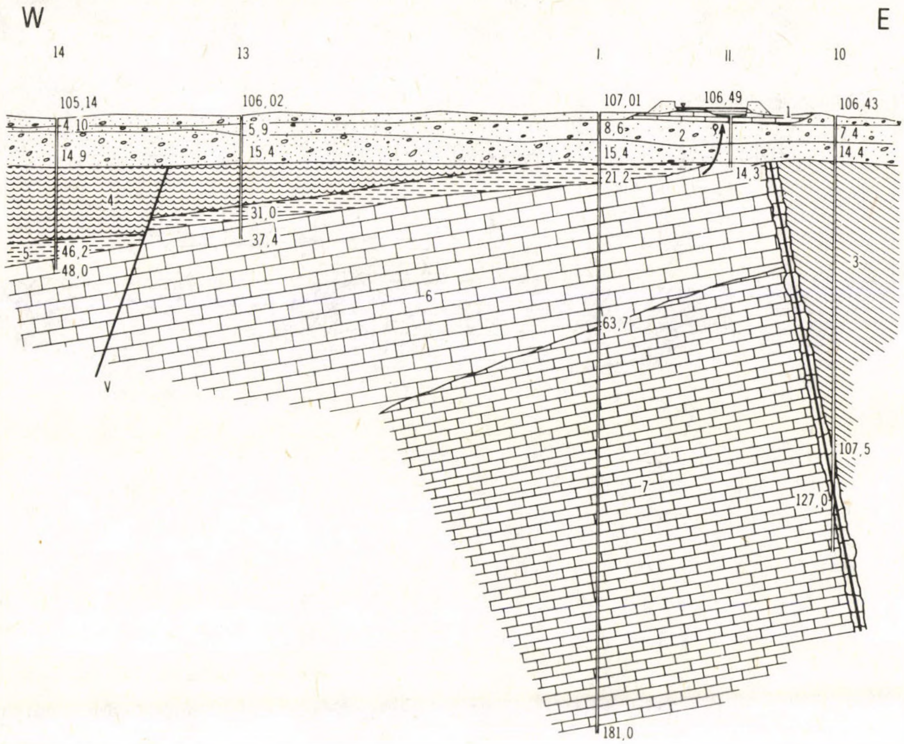


Fig. 3: The Római-fürdő Travertine (T2a - after SCHEUER, SCHWEITZER). 1 = loose, spongy travertine of lacustrine-marshy type; 2 = gravel with coarse sand, first flood-free terrace of the Danube (II/a - Würm); 3 = Middle Oligocene 'Kiscell Clay'; 4 = Lower Oligocene 'Tard Clay Marl'; 5 = Upper Eocene 'Buda Marl'; 6 = Upper Eocene limestone; 7 = Upper Triassic Dachstein Limestone

tetarata basin. The travertine was previously dated by Th/U method at 70 ka (PÉCSI, OSMOND, 1973), more recent datings (HENNIG et al. 1983; SCHWARCZ, SKOFLEK, 1982) indicated 100 ka and ESR analysis showed 127 ka (HENNIG et al., 1983). Consequently, the Tata Travertine may have formed in the (end of the) last interglacial, and early last glacial.

The same age can be estimated for the travertine in the foundations of the Radelkisz office building in the Bécsi út, Buda, earlier dated 70-75 ka (PÉCSI, OSMOND, 1973). Before Th/U analysis the travertines were thought to raise 4-5 m the surface of the first flood-free terrace (no II/b). Con-

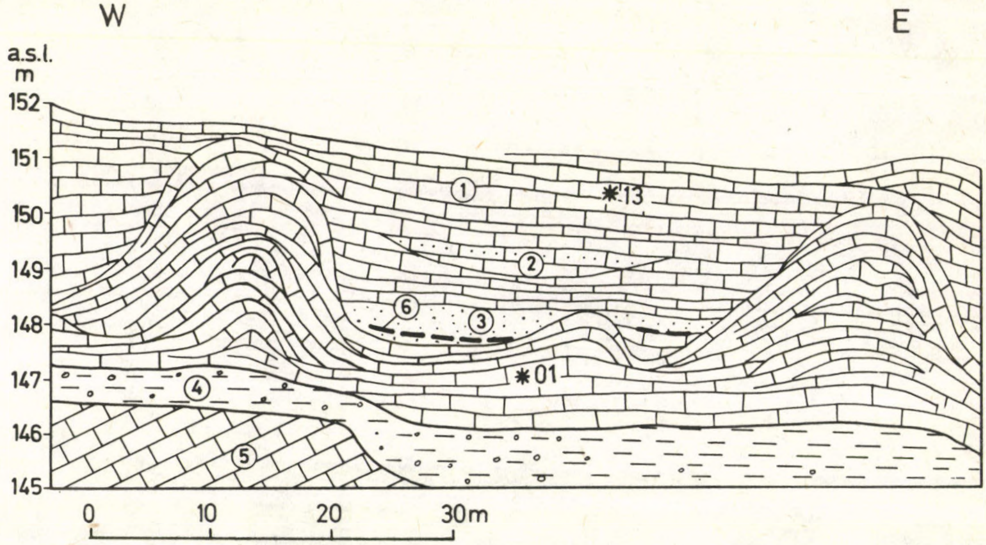


Fig. 4: The Tata Travertine (T2b), exposure of the travertine series on the II/b terrace of the Tata river and the geomorphological position of the Paleolithic site (after SCHEUER, SCHWEITZER).

1 = tetarata type travertine; 2 = calcareous silt accumulated in tetarata basin; 3 = eolian sand; 4 = terrace gravel and sand; 5 = Triassic limestone; 6 = Paleolithic site. The Th/U age of sample 01 is 101±10 ka BP, the Th/U age of sample 13 is 98±8 ka BP.

sequently, terrace material accumulated during the last interglacial or immediately before it (PÉCSI et al., 1985). N of Tata, the third flood-free terrace (no III) of the Danube (140 m above sea level) is overlain by travertine of only 135 ka Th/U age (PÉCSI, 1973; HENNIG et al., 1983). This lithostratigraphical unit is called Tata-Tóváros--Magdolna-puszta Travertine (T3a) and was deposited by the end of young Riss glacial and early last interglacial (Fig. 5). Thus, the Danubian terrace no III is of young Riss glacial age.

The Óbuda Travertine (T3b) overlies terrace on III no the Kiscell plateau, Budapest. This 10 m thin-bedded travertine was dated by the Th/U method (PÉCSI, OSMOND, 1973) at 175-190 ka (Fig. 6). It may represent early Riss glacial (KRETZOI, PÉCSI, 1982). Considerably older is the trav-

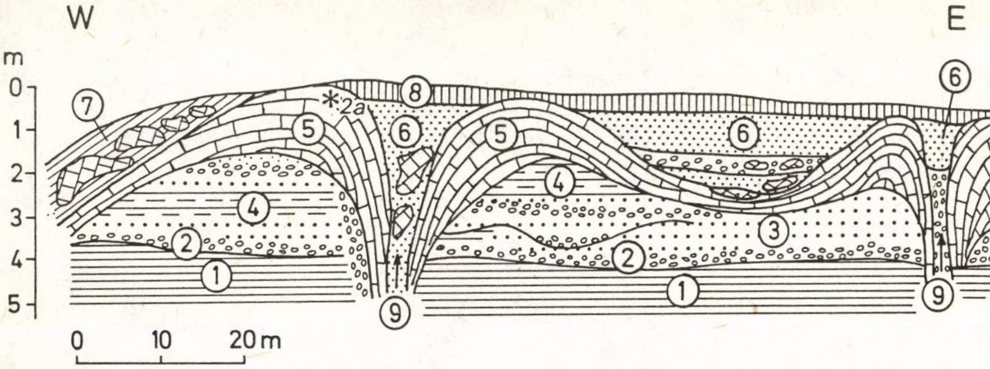


Fig. 5: The Tata-Tóváros -- Magdolna-puszta Travertine (T3a), spring cone type travertine on Danube terrace no III (after SCHEUER, SCHWEITZER).

1 = gray Pannonian clay; 2 = terrace gravel of the Danube; 3 = fluvial sand; 4 = silty sand; 5 = travertine; 6 = blown sand; 7 = slope deposit; 8 = recent chernozem; 9 = important spring; 2a = sampling site; Th/U age (analysis made in Florida) is 135 ka BP.

ertine on the flood-free terrace no III of the Tata river (Fig.7a), which has a Th/U age of 248 ka (HENNIG et al., 1983). This travertine is minimum one terrace younger and of lower geomorphological position than the Vértesszőlős Travertine with Lower Paleolithic culture.

The above suggest that two different phases of travertine formation (T3a and T3b) have to be identified. Thus, two terraces (IIIa and IIIb) can be distinguished. In the lower part of the Óbuda Travertine E. KROLOPP (1961) found mollusc remnants pointing to cool and humid environment. In the middle part molluscs favouring warm and dry, while in the top part those indicating cold and dry climate were found.

The 10-15 m Vértesszőlős Travertine unit (T4-T5) deposited on the alluvial fan terraces nos IV and V of the Tata river (PÉCSI, 1973). It contains some of the oldest (Lower Paleolithic) bone finds, tools, ovens and Upper Biharian (Mindel glacial) mammal remains (KRETZOI, VÉRTES, 1965).

The travertine with loess layers and sandy-detritic intercalations deposited from hot spring on the margin of an alluvial fan of prolonged accumulation (Figs. 7-8). Spring activity was interrupted several times and finally produced tetarata features between 350 and 400 ka BP (PÉCSI, 1973;

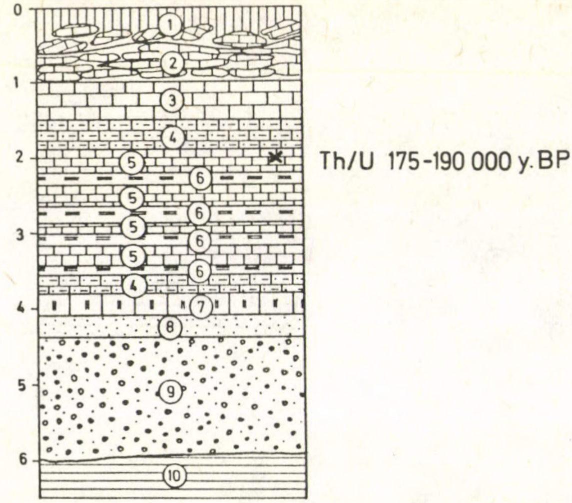


Fig. 6: The Óbuda Travertine (T3b). Profile of the travertine sequence deposited on terrace no IIIb of the Danube, Kiscell, Budapest (after SCHEUER, SCHWEITZER).

1 = recent rendzina soil; 2 = frost-decayed travertine; 3 = travertine in thick banks; 4 = calcareous silt; 5 = thin-layered travertine; 6 = fluvial silt and clay; 7 = boggy clay; 8-9 = fluvial sand, snady gravel; 10 = Middle Oligocene clay; Th/U age (samples dated by J.K. Osmond, Florida) is 175 ka BP.

HENNIG et al. 1983). The samples from the top of the travertine series, contaminated by calcareous silt and loess indicated much younger Th/U ages (SCHWARCZ, LANTHAM, 1984); they are unacceptable datings. Similarly, the dating of travertine samples contaminated by calcareous material also resulted in incorrectly young age data (HENNIG et al., 1983). Reliable data can only be obtained from compact travertine layers (Fig. 7).

The Budavár Travertine (Fig. 9) contains Mindel-Riss intergalcial fauna. The absolute age of this travertine (T4) was determined by the Th/U method at 380 ka (HENNIG et al., 1983). From the karstic hollows of the travertine finds representing the Castellum faunal phase were recovered. In this base the gravel of the terrace IV of the Ördögárok stream is found. The accumulation of the Buda Castle Travertine series continued to the penultimate interglacial (KROLOPP et al., 1976). S of Almásneszmély the no V terrace of the Danube (175-185 m above sea level) is overlain by the Vöröskő Travertine (T5 - Fig. 10). It has a thickness of

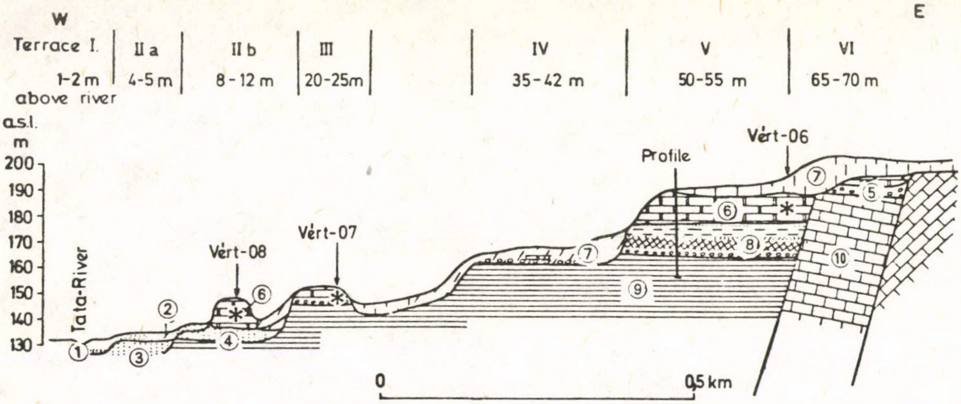


Fig.7a: Detailed geomorphological profile of the Tata river valley with the Paleolithic site at Vértesszőlős (after PÉCSI)

1 = flood-plain; 2 = colluvium; 3 = sandy gravels of the first flood-free terrace; 4 = fluvial sand (tIIb); 5 = thin gravel beds (tIIb to VI); 6 = travertine cap on terraces III to VI; 7 = loess, slope loess and slope deposit; 8 = alluvial fan and terrace series with red clay; 9 = Oligocene clay, sandy clay and gravel; 10 = Triassic limestone; ⁎ = Paleolithic sites

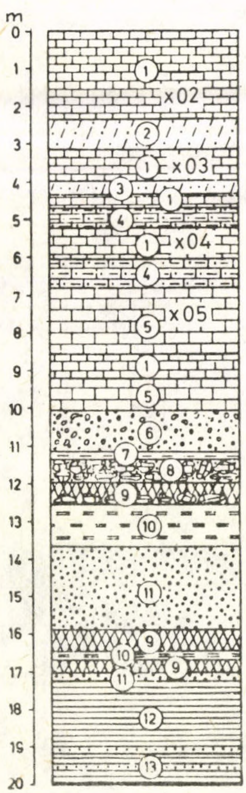


Fig.7b: The Vértesszőlős Travertine (T4-5)

1 = loose, thin-layered travertine; 2 = stratified loess; 3 = fine sand with calcareous silt; 4 = calcareous silt; 5 = compact, thick-banked travertine; 6 = sandy pebble; 7 = ochre clay; 8 = local alluvial fan gravel, coarse limestone gravel and pebble; 9 = red clay paleosols; 10 = variegated terrestrial clay; 11 = fluvial sandy pebble; 12 = Oligocene clay; 13 = Oligocene clayey sand; ⁎ = Lower Paleolithic cultural layers; 02,03,04 and 05 = sample sites for Th/U and ESR datings

Th/U and ESR datings after G.J. HENNIG et al. (1983):

Th/U	ESR
128^{+20}_{-12} ka^{x}	127 ± 13 ka^{x}
217^{+40}_{-28} ka^{x}	245 ± 25 ka^{x}
325^{+00}_{-60} ka^{x}	172 ± 17 ka
350 ka	333 ± 17 ka

^{x}contaminated samples
Th/U dating after J.K.OSMOND (1973), two samples analyzed on 3 occasions: 350 ka BP.


Lake Tata
Tata
Vértesszőlős
0 500 1000 m
1 2 3 4 5 6 7 8 9 10 11 12 13 14 15
16 17 18 19 20 21 22 23 24 25 26 27 29 30 31

Fig. 8: Geomorphological map of the Vértesszőlős-Tata area (after PÉCSI-SCHWEITZER).

1 = Lower Tertiary planation surfaces on Mesozoic horsts in the Gerecse Mountains; 2 = Mesozoic horst in threshold position at Tata; 3 = remnants of Upper Miocene raised beaches; 4 = pediment and margin (locally two levels); 5 = terrace no VI; 6 = terrace no V; 7 = terrace no IV; 8 = terrace no III; 9 = terrace no II/b; 10 = terrace no II/a; 11 = higher flood-plain level, terrace no I; 12 = lower flood-plain level; 13 = seasonally waterlogged flood-plain; 14 = terraces covered by travertine; 15 = natural and artificial lakes; 16 = gully; 17 = karst erosional valley; 18 = erosional-derasional valley; 21 = doline; 22 = col; 23 = settlement; 24 = strip mine; 25 = steep slope; 26 = gentle slope; 27 = lapiés slope; 28 = alluvial fan; 29 = Vértesszőlős Paleolithic site; 30 = Tata Paleolithic site; 31 = Kenderhegy Paleolithic site

25 m, the top larger part is strongly sandy limestone or sand with calcareous cement. Paleomagnetic dating (by P. MÁRTON) placed the beginning of travertine accumulation to times prior to the Jaramillo event. Thus, it belongs to the upper part of the Lower Biharian stage (0.7-0.9 ma BP). The geomorphological position suggests that the travertine here directly overlies the no V Danubian terrace (PÉCSI, 1959; PÉCSI-SCHEUER-SCHWEITZER 1982). The upper travertine has been removed locally (Fig. 11).

The Ürömhegy Travertine lies at 190-210 altitude, N of Budapest. It contains Biharian fauna (JÁNOSSY, 1979). The mammal and mollusc faunae (KROLOPP, 1961) and the geomorphological position (Fig. 12) equally support its classifying as the upper part of Lower Pleistocene. In its base terrace sand is found.

The terrace no V and the travertine no T5 also seem to represent two lithostratigraphical units of slightly different age. The Szomód Travertine on the Leshegy, Gerecse Mountains (230 m above sea level) is similar to the Ürömhegy one but somewhat older. Archidiskodon meridionalis meridionalis teeth were found. In the base Danubian gravel of considerable (25-30) thickness is disclosed.

The terraces nos VI and VII of the Danube in the environs of Dunaalmás are mantled by 25-30 m tetarata type travertine. Here too this hard rock

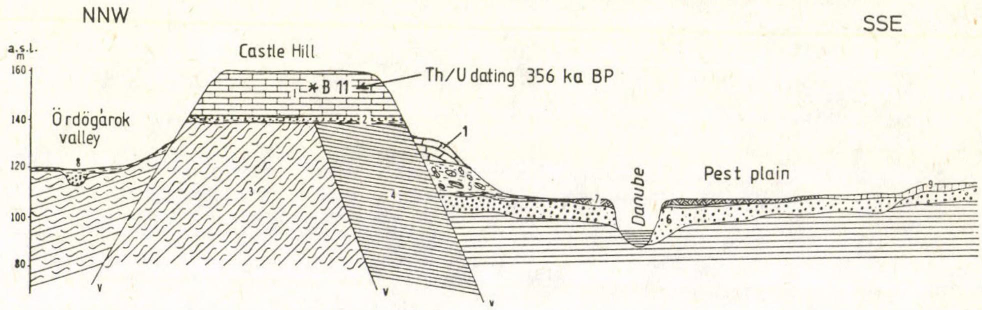


Fig. 9: The Budavár Travertine (T4). Geological profile across the Buda Castle Hill (after SCHEUER - SCHWEITZER).
1 = thick-banked travertine of lacustine-marshy type with intercalated loess or paleosols; 2 = terrace deposits of the Ördögárok stream; 3 = Upper Eocene calcareous marl; 4 = Middle Oligocene clay; 5 = slope debris; 6 = Danubian terrace deposits; 7 = artificial fill; 8 = recent deposits of the Ördögárok stream; 9 = loessy sand; ⁎ = Th/ U dating: 350 ka BP (HENNIG et al., 1983)

has preserved the terraces from denudation. The two members of the D u n a - a l m á s T r a v e r t i n e closely resemble in structure (Fig. 13). In both thick banded travertine series have 4-5 intercalations of loess and loessy sand and a ochre-red paleosol, indicating the interuption of spring activity in the tetarata basin. From the ca 1 m red paleosol intercalated between the layers of the higher-lying T7 travertine (Fig. 14) K i s l á n - f i a n f a u n a was described by D. JÁNOSSY (1979). A similar red soil occurs at 15-20 m lower intercalated in the travertine T6. Therefore, the travertines T7 and T6 also belong to the lowermost Pleistocene (PÉCSI-SCHEUER-SCHWEITZER 1982; SCHEUER-SCHWEITZER 1974).

The Upper Villányian stage (the Kislángian) dates from ca 2 ma to 1.4 ma, the lower boundary coinciding with the Plio-Pleistocene boundary. In our opinion, most of the Kislángian substage belongs to the Lower Pleistocene as well as the Lower Biharian substage (KRETZOI- PÉCSI, 1982; SCHEUER-SCHWEITZER 1983; JÁNOSSY 1979).

The presence of a red soil in the travertines T6 and T7 certainly indicate a considerable interruption in travertine deposition. According to

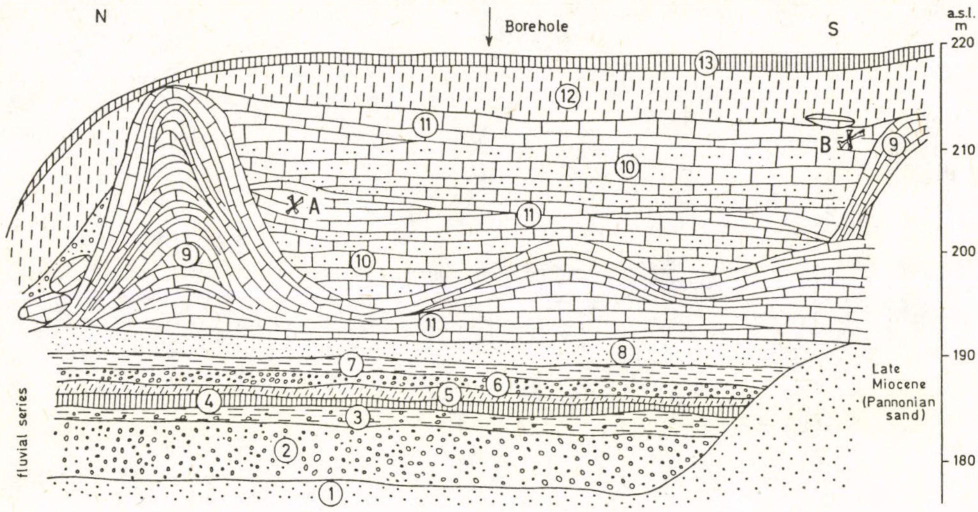


Fig. 10: The Vöröskő Travertine (T5), S of Almásneszmély (Dunaalmás - after SCHEUER-SCHWEITZER).

1 = medium grained, micaceous, yellow sand; 2 = sandy, spherical pebble, material of terrace no V of the Danube; 3 = micaceous, gravel (1-2 cm diameter) sand; 4 = humic sand; 5 = light yellow clay with gravels of travertine and quartz, divided by 1-5 cm loose travertine layer; 6 = travertine and quartz gravels in yellowish gray silt (ravine deposit); 7 = yellowish gray, calcareous, sandy silt; 8 = calcareous sand; 9 = compact, sugar-textured travertine, part of a tetarata barrier; 10 = fluvial sand with calcareous silt; 11 = compact, banked travertine; 12 = loessy sand, loess with frost shattered travertine at base; 13 = recent soil. A = Clemmys méhelyi Kormos (= Emys orbicularis L.); Megaloceros sp.; B = Archidiskodon meridionalis (planifrons) finds. Reverse polarity by paleomagnetic investigation.

the paleomagnetic investigations by P. MÁRTON and M.A. PEVZNER (Fig. 14), the 24 m of the travertine series T7 shows reverse polarity. Only one sample from the bottom layer was of normal polarity. Interpreting the results, the loess layers and the ochre-red paleosol intercalated in the travertine T7 are dated 1.4-1.7 ma BP. The Lower Pleistocene (Kislángian) travertines T6

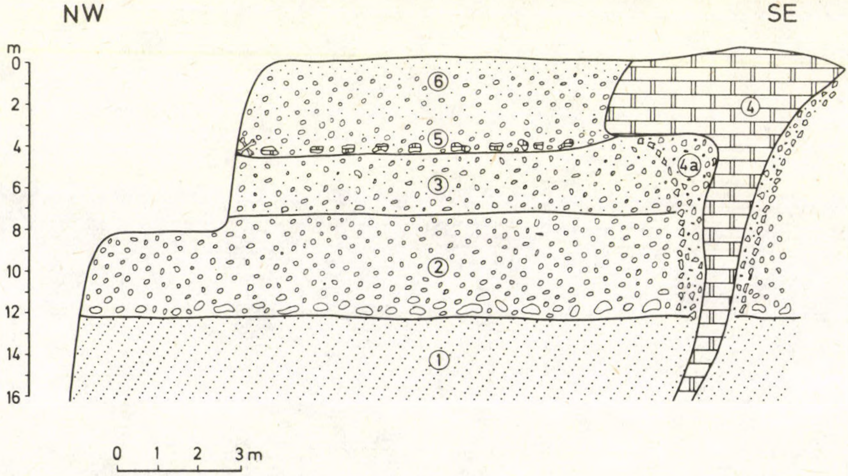


Fig. 11: The Dunaalmás Gravel. Profile of alluvial fan terrace no V of the Danube S of Almásneszmély, in the exposure of the gravel pit at the '181 m' triangulation point (after PÉCSI).

1 = Upper Pannonian cross-bedded sand; 2 = mostly coarse gravel with boulders (20–40 cm diameter) gneiss, granite and metamorphic rocks (not deltaic, but alluvial fan channel deposition); 3 = coarse and medium grained gravels with sand intercalation; 4 = spring vent filled by travertine; the travertine body may have been eroded; 4a = gravels oriented parallel to the travertine body; 5 = coarse and medium grained gravel with sand layers, slightly spherical travertine debris at base; 6 = medium grained gravel (average diameter: 2–5 cm), upward transition to sand. The exposure lies on hillslope, at the intersection of layer 3 and slope surface ventifacts are common.

and T7 are both of considerable (more than 30 m) thickness. Regarding their origin, they are of tetarata type overlying gravel terraces. These Lower Pleistocene travertines can be regarded key sections and lithostratigraphical units under the name Dunaalmás Travertine series.

NEOGENE TRAVERTINES AND GEOMORPHOLOGICAL SURFACES

In addition to the Quaternary travertines, in the Transdanubian Mountains (particularly in the Gerecse and the Buda Mountains) 4-5 generations

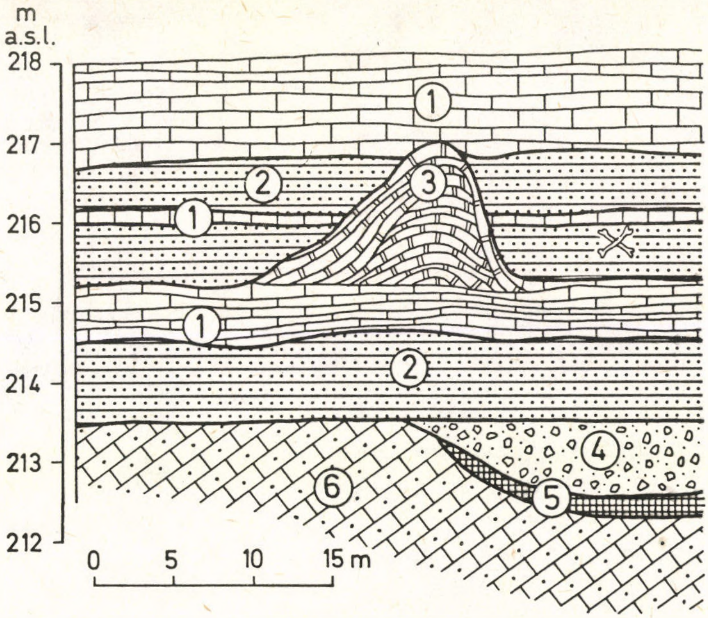


Fig. 12: The Ürömhegy Travertine (T5), N of Budapest (after SCHEUER-SCHWEITZER)

1 = compact, thick-banked travertine; 2 = calcareous silt; 3 = compact, thin-bedded, travertine, tetarata barrier; = faunal site (Mimoys savini Hinton, Archidiskodon, Trogontherii cromerensis Depéret et Majet); 4 = dolomite debris; 5 = red clay; 6 = Upper Triassic dolomite

of Neogene travertine also occur on higher geomorphological surfaces. The position of travertines on the geomorphological surfaces on the margin of the Gerecse along the Danube is shown in Fig. 15.

The Süttő Travertine (T8) lies on the lower pediment of the Gerecse looking on the Danube (at 240-280 m altitude - Fig. 16-17.). The compact travertine is quarried for ornamental stone. It contains Upper Pliocene Csarnótan and Ruscinian vertebrate fauna (JÁNOSSY, 1979; SCHEUER-SCHWEITZER 1983). Total thickness is 25-30 m. In the cover of the thick-banked travertine sporadical gravel, red soil, some old loess and young loess are found (Fig. 18).

The N part of the extensive travertine lies ca 20 m lower in the Diósvölgy quarry overlying a formation of pebbles and sand (presumably the

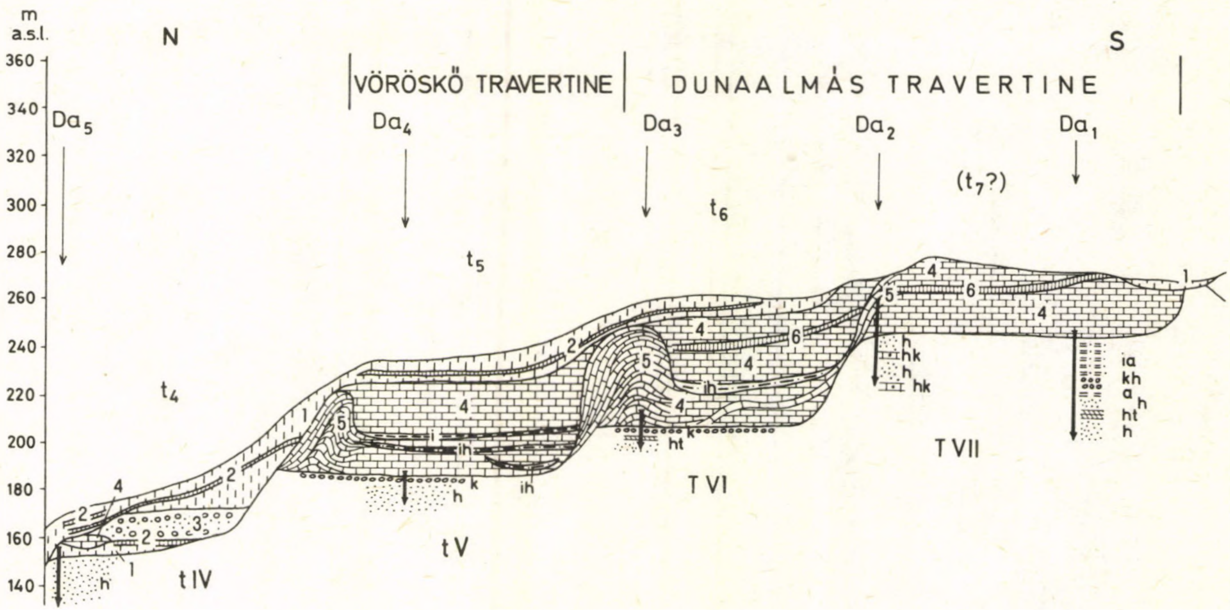


Fig. 13: The Dunaalmás Travertine (T6–T7). Profile of the Danube terraces nos IV–VII and the overlying travertine sequences (T4–T7) based on exposure and borehole data (after PÉCSI-SCHEUER-SCHWEITZER 1980). 1 = loess and slope loess; 2 = paleosols in loess; 3 = terrace gravel; 4 = travertine; 5 = terrace tetarata barriers; 6 = paleosol in travertine, Da_1–Da_5 = borehole sites, tIV – tVII = terraces, a = clay; ia = muddy clay, ih = muddy sand, h = sand, kh = gravelly sand, ht = hydromorphous soil, hk = sandstone, k = gravel

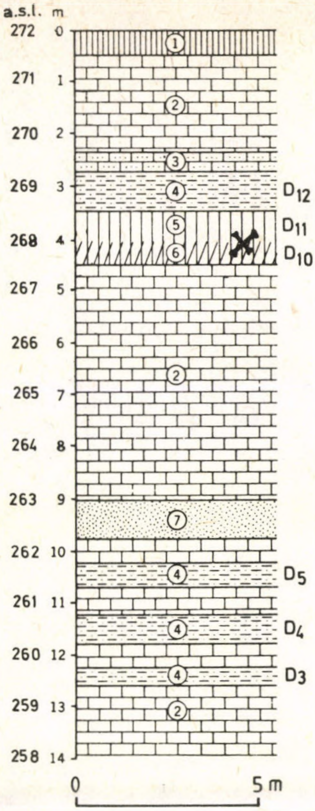


Fig. 14: The Dunaalmás Travertine (T7). Profile drawn by F. SCHWEITZER, paleomagnetic investigations by M.A. PEVZNER and fauna identification by D. JÁNOSSY (1979)
1 = recent rendzina soil; 2 = travertine; 3 = calcareous silt; 4 = loess with fine sand; 5–6 = double paleosol, no 6 is Lower Pleistocene ochre–red soil of Kislángian type with rich micro- and macrofauna; 7 = silty sand, D_3–D_{12}= samples for paleomagnetic analysis (all are of reverse polarity)

oldest - no VII or VIII? - terrace of the Danube).

The Kőhegy Travertine (T9) is located on the margin of the Gerecse (Fig. 15) on a raised beach at 290 m above sea level. Compact travertine is ca 30 m thick with numerous bivalves belonging to the Unio wetzleri horizon in the top part. This mollusc fauna is referred to the Upper Pontian stage of the Miocene (SCHWEITZER -SCHEUER, 1983). Of similar age are the Nagyvázsony Travertine in the Bakony Mountains (JÁMBOR, 1980) and the Várpalota Kálvária Hill Travertine (BARTHA, 1959).

The Kőpite Travertine (T10), deposited from a large and several smaller hot spring vents, forms a distinct geomorphological surface. An Anancus arvernensis tooth has been recovered from it (JÁNOSSY, 1979). The travertine body of frustum cone shape may have preserved a higher, Pliocene, pediment and Miocene (Upper Pontian) delta gravel from denudation. In the base Pontian pebble and coarser gravel both of deltaic bedding are deposited over Pannonian coarse sand of considerable thickness as exposed in the neighbouring gravel pits (Fig. 19). In the top part of pebble thin basalt tuff strings are also observed. The two gravel series are divided by spherical travertine debris.

The travertine blocks embedded in the gravel base of the Kőpite Travertine were reworked from the nearby, 30-40 m thick, Dunaszentmiklós

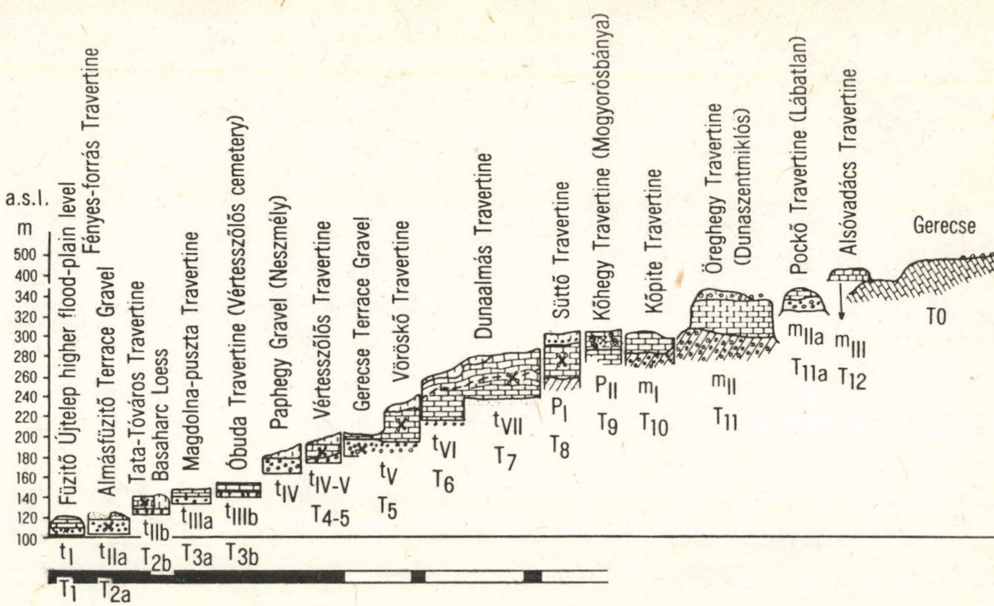


Fig. 15: Geomorphological surfaces and travertine horizons in the Gerecse foreland (PÉCSI-SCHWEITZER-SCHEUER 1986)
tI-tVII = river terraces usually covered by travertines (T1-T7) and loess; PI-PII = Pliocene pediment surfaces covered by travertines (T8-T9); mI-mIII = Upper Pannonian (Upper Miocene) raised beaches covered by travertines (T10-T12); T0 = Paleogene-Mesozoic planation surface sculptured by Oligocene-Miocene pedimentation with sporadic gravels.

T r a v e r t i n e (on Öreghegy, at 330 m above sea level). Although this forms an extensive plateau, it has not been dated yet (T11 on the comprehensive profile of geomorphological surfaces in the side of the Gerecse - Fig. 15).

Some additional, higher (at 330-350 m altitude) occurrences of travertine (T11 and T12) on the N margin of the Gerecse overlie Upper Miocene raised beaches and delta gravels. They have not been dated precisely.

The highest Neogene raised beaches with travertine mantles (T11 and T12) containing fauna suitable for dating (Fig. 21) occur in the E of the Buda Mountains. The position of these geomorphological surfaces (Fig. 20) points to more intensive uplift in the E part of the Buda Mountains than in its W part and S foreland. In the latter zone Miocene (Sarmatian) limestones

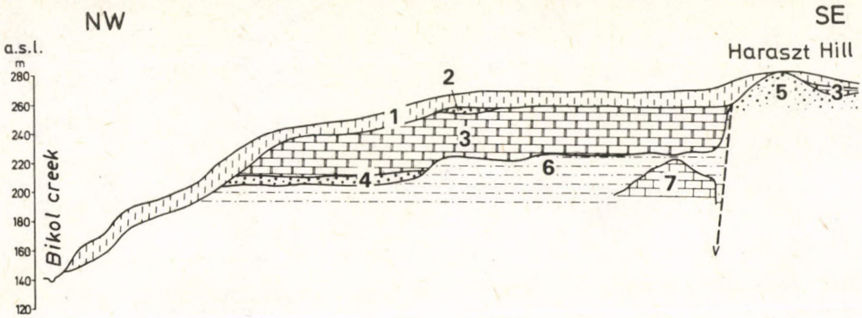


Fig. 16: The Süttő Travertine (T8). Sketch profile based on borehole data of quarry (after PÉCSI)

1 = young loess with paleosols, below the paleosol MB (last interglacial) locally remnants of old loess; 2 = sporadic medium grained (2-3 cm diameter) spherical quartz gravel, in patches covered by clay weathering layers containing cray fish (Potamon sp.) and bone fragments (Upper Pliocene Csarnótan faunal phase); 4 = pebble (of uncertain age, Upper Pliocene? lowermost Pleistocene?); 5 = sandy detritic pebble (Upper/most/ Pliocene, Pontian); 6 = clay, clayey sand series (Upper Pontian); 7 = Jurassic limestone breccia

(also littoral formations) have elevations of 250-350 m above sea level, while in the E-Buda Mountains younger (Pontian) raised beaches and the overlying travertines attain altitudes of 400-500 m.

The Széchenyi-hegy Travertine (T10) has 10-15 m average thickness and mantles the 410-430 m Upper Pontian raised beach (littoral sands - Fig. 22).

The predominatly terrestrial fauna collected from the thickbanded travertine with thermal karst caverns (KRETZOI, 1978) point to Upper Miocene (Sümegian) age corresponding to the Upper Pannonian (sensu strictu) Baltavárian formations. The investigations by J. OPDYKE showed that the upper part of the travertine has reverse polarity.

The Szabadság-hegy Travertine (T11) is a ca 20 - 25 m, compact, thick-banded series in 30-40 m higher position than the former (at 440-470 m). The base is (Lower) Pannonian (sensu strictu) raised beach (Fig. 23), sandy pebble from which Upper Miocene rhinoceros (Aceratherium incisivum KAUP sp.) was recovered. In the opinion of M. KRETZOI this find

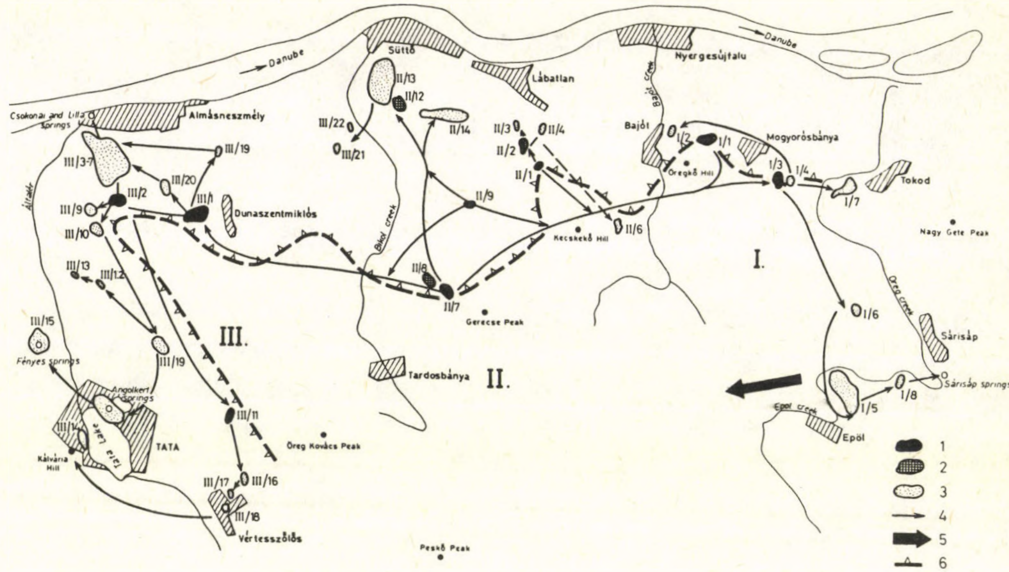


Fig. 17: Principle travertine occurrences in the Gerecse Mts. - (after SCHEUER - SCHWEITZER)

1 = Upper Pontian (Pannonian) springs and travertine; 2 = Upper Pliocene springs and travertines; 3 = Quaternary springs and travertines; 4 = direction of spring displacement; 5 = direction of Upper Pleistocene spring displacement; 6 = presumed Upper Pontian (Pannonian) shoreline; I = E-Gerecse, II = Central Gerecse, III = W-Gerecse. - I/3 = Kőhegy travertine, II/12, II/13 = Süttő Travertine, III/1 = Dunaszentmiklós Öreg-hegy Travertine, III/2 = Kőpite Travertine, III/3 - III/7 = Dunaalmás (Almásneszmély) and Vöröskő Travertines, III/12-13 = Tata-Óváros-Magdolna-puszta Travertine, III/14 = Tata Travertine, III/15 = Fényes-forrás Travertine, III/16-18 = Vértesszőlős Travertine, III/19-20 = Meleges-hegy, III/21-22 = Fekete-hegy-Süttő Pörösföldek

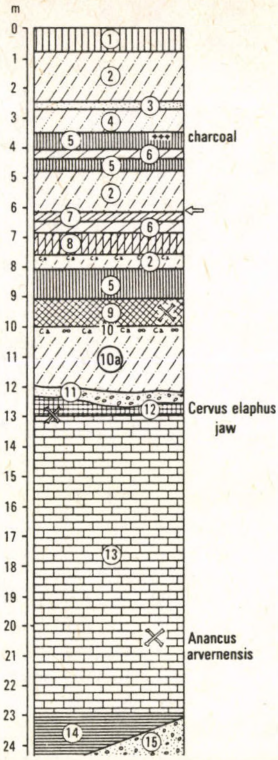


Fig. 18: Stratigraphy of the Süttő Travertine and the overlying loess (after PÉCSI-SCHWEITZER)

1 = recent chernozem; 2 = stratified slope loess with fine sand; 3 = fine sand; 4 = loessy fines sand, dell fill; 5 = 'humus carbonate' steppe soil; 6 = clayey slope loess; 7 = semipedolite; 8 = fossil chernozem; 9 = fossil reddish brown forest soil; 10 = Ca accumulation horizon; 10a = old loess with fine sand; 11 = gravel and sand; 12 = red clay; 13 = compact travertine of lacustrine-marshy type with homogeneous crystalline structure, Potamon sp. and Anancus arvernensis finds; 14 = Upper Pontian (Pannonian) clay layers; 15 = Upper Pontian (Pannonian) pebble and sand

represents the Lower Pannonian (Eppelsheimian faunal stage). By the Tapiriscus (KRETZOI, 1978) and Hipparion finds (JÁNOSSY, 1979) recovered from the travertine with Melanopsis on the raised beach, Gy. SCHEUER and F. SCHWEITZER, 1973, 1983 correlated the earlier Upper Pannonian (Csákvárian faunal stage).

Ca 20-30 m higher than the former, there is another lithologically distinct travertine occurrence on the Szabadság-hegy (at 475-490 m above sea level). It is a thick-banded, compact, brownish grey series with a slightly bituminous smell. Its base has not been revealed and no fauna indicating age has been found. It is assumed to be older than the Szabadság-hegy Travertine. From lithological and geomorphological evidenceGy. SCHEUER and F. SCHWEITZER (1973, 1983) propose it to mark as a separate lithostratigraphical unit under the name of H á r m a s k ú t t e t ő T r a v e r t i n e (T12).

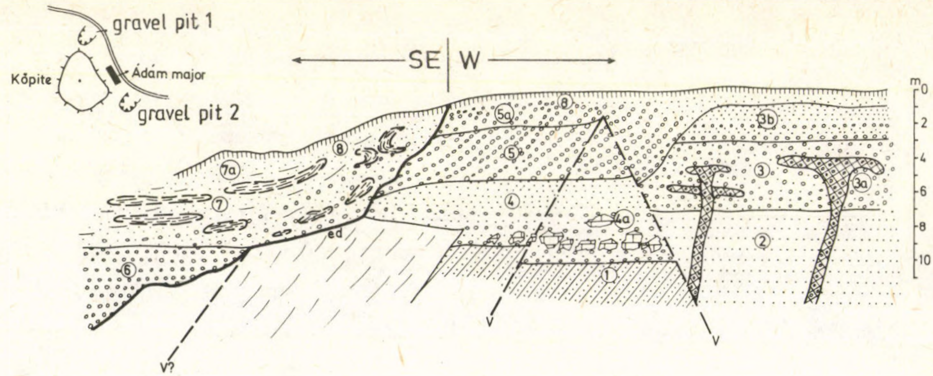


Fig. 19a: Sketch profile of the gravel pit no 2 on the E slope of the Kőpite (after PÉCSI)

1 = Pannonian sandy clay; 2 = coarse white sand, plane bedded, locally cross-bedded; 3 = sandy quartz pebble, many small (1 cm diameter) black (basalt?) pebbles; 3/a = spring vent with sand and gravel cemented into calcareous conglomerate; dragging of gravel and sand along the vent; 3/b = quartz sand with infrequent quartz pebble interbeddings; 4 = alternating white quartz gravel and pebble; 4/a = travertine blocks (0.3 to 1 m diameter) in white quartz sand matrix; 5 = delta-bedded fine and medium grained spherical gravel; 5/a = plane-bedded gravel, mixed and non-stratified in the upper part; 6 = predominantly coarse gravel, not deltaic, composed mostly of quartzite, but significant numbers of granite, gneiss, lidite, metamorphic and travertine gravels; 7 = white sand with silty, calcareous and clay layers; 7/a= grayish sand; 8 = humus with slope debris, v = fault, proved; v?= hypothetical fault, ed = striking erosional unconformity

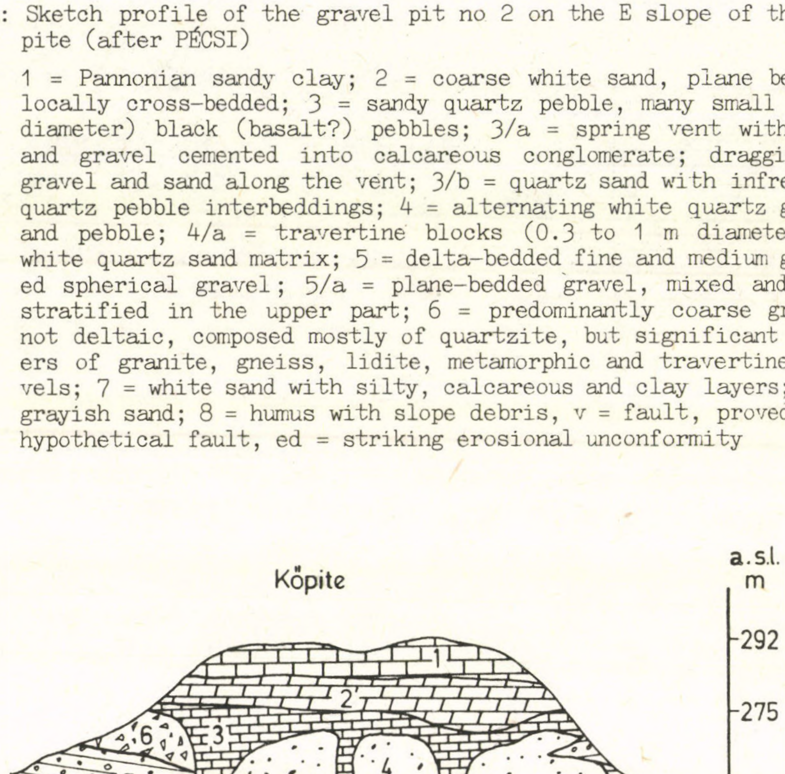


Fig. 19b: The huge cones of the Kőpite Travertine (after PÉCSI)

1 = ochre-yellow compact travertine; 2 = light compact travertine; 3 = dyke-form travertine; 4 = Pannonian coarse sand; 5 = Pannonian delta-bedded gravel; 6 = coarse slope deposit

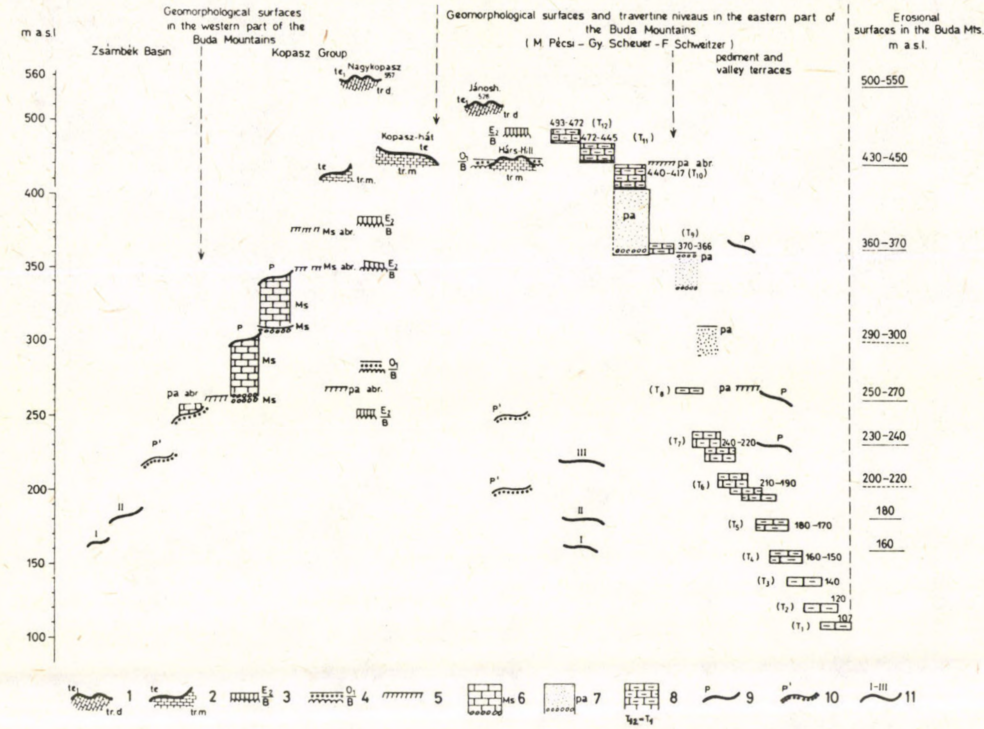
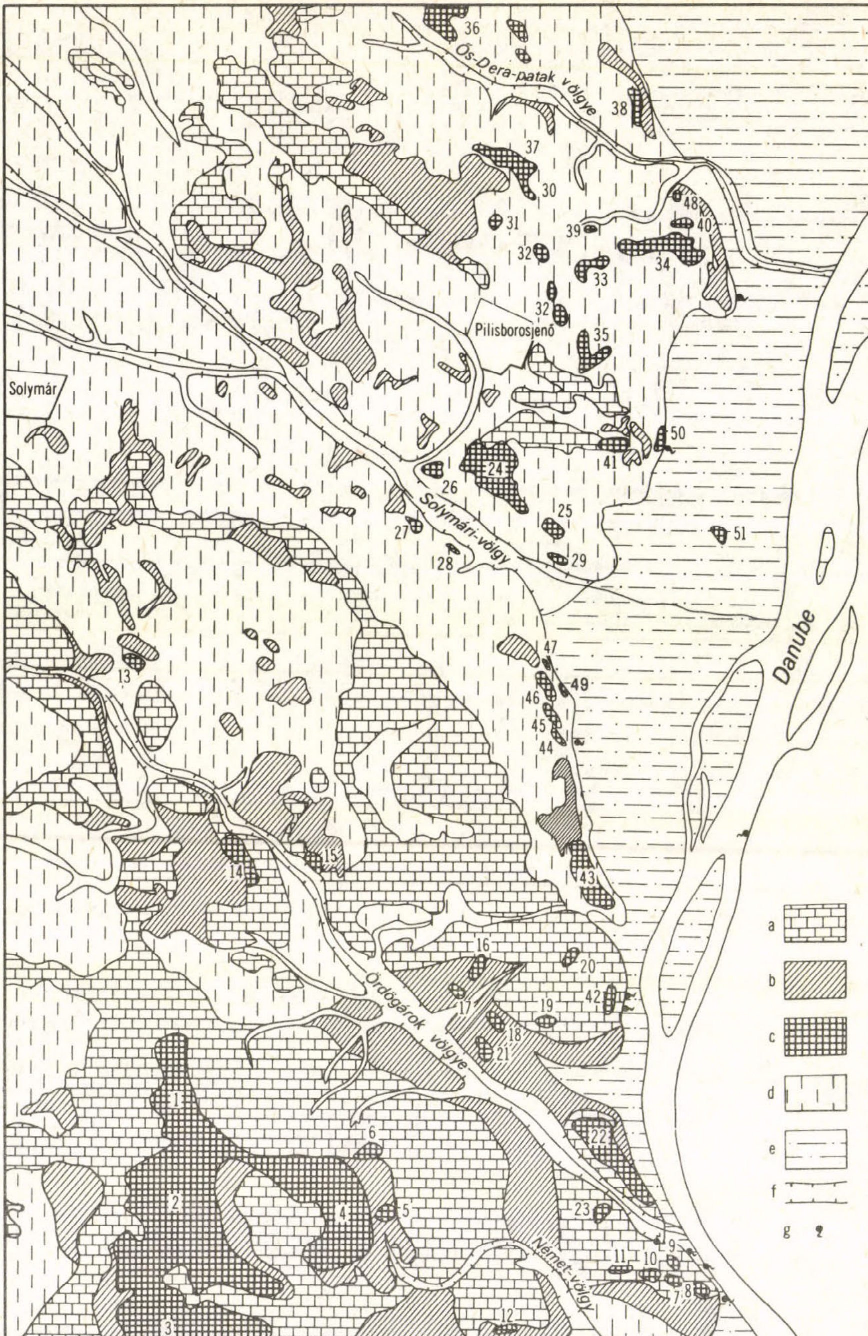


Fig. 20: Geomorphological surfaces in the Buda Mts (after PÉCSI-SCHEUER-SCHWEITZER, 1979 - based on data by PÉCSI, 1963, 1975; SCHEUER-SCHWEITZER, 1974; WEIN, 1977)

1 = exhumed Mesozoic peneplain in summit position (te_1) on Upper Triassic dolomite (tr.d.), 2 = remnants of exhumed Mesozoic peneplain (te) on Upper Triassic Dachstein Limestone (tr.m); 3-4 = buried Mesozoic peneplain, remnants of tropical karst and bauxite under Eocene limestone (E_2/B) or under Oligocene sandstone (O_1/B); 5 = raised beach; 6 = Miocene (Sarmatian) gravel and coarse grained limestone (Ms/s); 7 = Pannonian (pa) gravel, sand and clay; 8 = travertine horizons (T12-T1); 9 = Pliocene pediment (P) on solid rock; 10 = Pliocene pediment on unconsolidated deposits (pl); 11 = Pleistocene derasion terraces, gentle slope segments on unconsolidated deposits



Ős-Dera-patak völgye
Pilisborosjenő
Solymár
Solymári-völgy
Danube
Ördögárok völgye
Német-völgy
a
b
c
d
e
f
g

Fig. 21: Occurrences of travertine in the Buda-Pilis Mts (after SCHEUER-SCHWEITZER)

a = Upper Triassic limestone and dolomite, b = Paleogene formations, c = occurrences of travertine on surfaces of various altitude, d = loess and slope loess, e = alluvium, f = erosional valley, g = existing lukewarm karst springs; 1-51 = number in travertine inventory;

1-2 = Szabadság-hegy: 1 = Hármaskút-tető; 2 = Observatory, Pioneer's Camp; 3 = Budaörsi-hegy, Kakukk-hegy; 4-6 = Széchenyi-hegy: 5 = Felhő u. 7; 6 = Alkony utca; 7-11 = Gellért-hegy: 7 = Jubileumi park; 8 = Liberation Monument; 9 = Számadó u. 7; 10 = Kelenhegyi út; 11 = Somlyói út; 12 = Sas-hegy; 13 = Máriaremete; 14 = Hűvösvölgy, Nyéki út; 15 = Hűvösvölgy, Kondor út; 16-19 = Rózsadomb: 16 = Törökvészi út; 17 = Lepke utca; 18 = Vérhalom; 19 = Bimbó út; 20 = Szemlő-hegy; 21 = Viticultural Research Institute; 22 = Castle Hill; 23 = Nap-hegy; 24 = Üröm-hegy upper; 25 = Arany-hegy upper; 26 = Üröm-hegy lower; 27 = Csúcshegy dűlő upper; 28 = Csúcshegy dűlő lower; 29 = Arany-hegy lower; 30 = Harapovácsi upper; 31 = Monalovác-hegy, S slope; 32 = environs of Puszta-hegy; 33 = Kálvária-tető upper; 34 = Ezüst-hegy lower; 41 = Péter-hegy; 42 = Rózsadomb, Apostol u. 15-17; 43 = Kiscell plateau; 44 = Farkastorki út; 45 = Farkastorki lejtő; 46 = Labanc köz upper; Labanc köz lower; 48 = Budakalász; 49 = Bécsi út; 50 = Csillaghegy swimming pool; 51 = Római-fürdő (Roman baths)

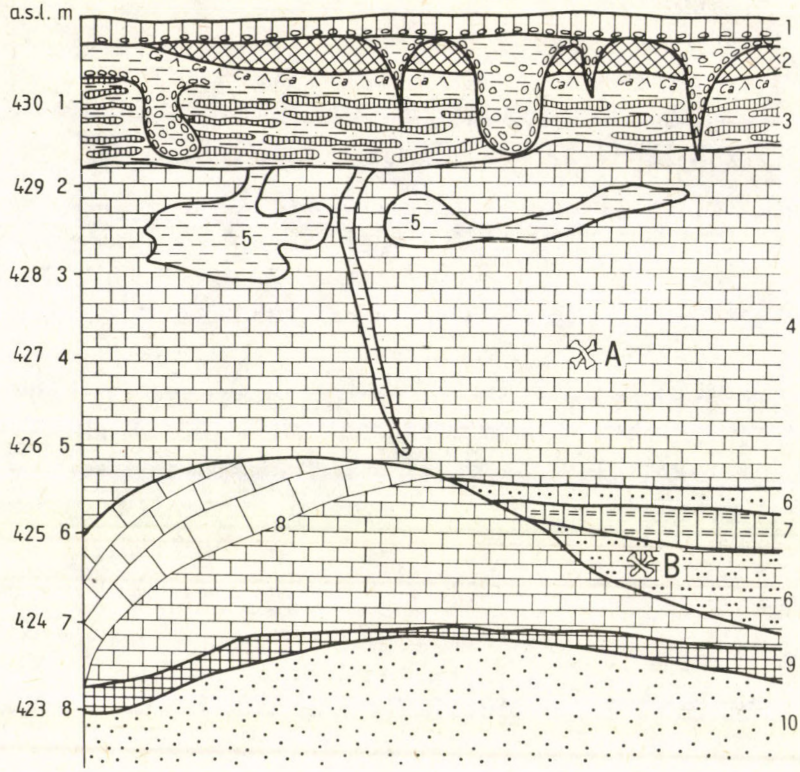


Fig. 22: The Széchenyi-hegy Travertine (T10), Buda Mts, 420 m above sea level (after SCHEUER - SCHWEITZER, 1974)

1 = recent rendzina soil; 2 = ochre-brown forest soil with frost phenomena on its surface; 3 = frost-riven travertine; 4 = compact crystalline travertine with thermal hollows; 5 = reddish brown clay with travertine detritus; 6-7 = calcareous silt (6) in the tetarata basin and grayish hydromorphous soil (7); 8 = margin of the tetarata basin; 9 7 bright red clay; 10 = Upper Pontian yellow micaceous sand, A = Giraffida sp., Tapiriscus sp. finds (of Sümegian age), B = Parapodemus sp., Gerbillida sp., Ochatonida sp. (Upper Miocene)

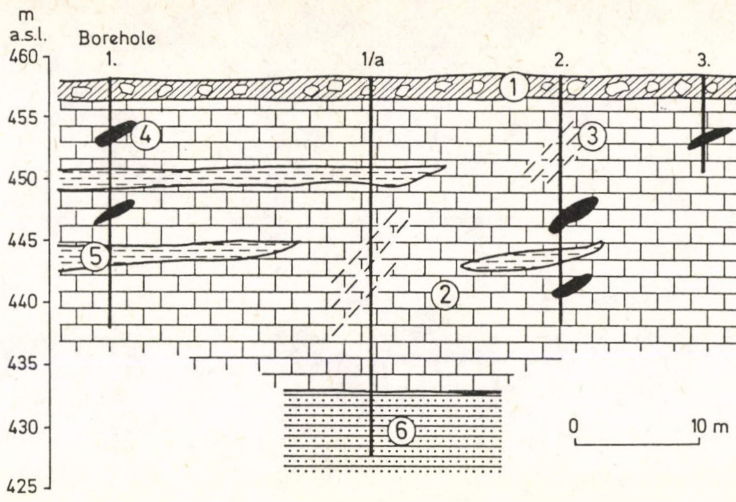


Fig. 23: The Szabadság-hegy Travertine (T11 - after SCHEUER - SCHWEITZER)
1 = rendzina; 2 = compact, light brown travertine with Melanopsis, Hipparion and Tapiriscus fauna; 3 = karst hollow with reddish brown clay fill; 4 = karstic hollow; 5 = calcareous silt; 6 = Upper Miocene (Pannonian) sand with banks of sandstone, locally with quartz pebble and Aceraterium incisivum fauna

CONCLUSION

In the paper a brief outline was given of the young Cainozoic denudation chronology of the Transdanubian Mountains resulting from several decades of research applying multidisciplinary methods in systemizing chronologically the geomorphological surfaces. They provide the basis for future research and comparisons.

The 12-15 geomorphological surfaces described in the paper represent the individual stages of the geomorphic evolution during the last 10 Ma. In reality, more geomorphological surfaces may have formed, but some of them (mainly the older ones) were destroyed and some could not be distinguished as yet. However, the systematization of the younger (Pleistocence) surfaces is a good starting point for the identification and explanation of the representative stages in geomorphic evolution, for theoretical work and for drawing conclusions for practical purposes (such as neotectonics or stone-quarrying).

Table 1: Late Cainozoic biostratigraphy and geomorphological surfaces (after KRETZOI 1927–1983 and PÉCSI 1983)

Ma	Epoch	Stage	Substage	Stratotype localities	Travertine	Terraces	Pediment, foothill surface	Loess, paleosols fluvial, lacustr. sed.	Localities and notes
0.01		Holocene			N°1	N°I		flood plain sed.	
0.2	PLEISTOCENE	PILISIUM	Szántóium	Pilisszántó	N°2	N°IIa		Young loess with 5 paleosols	Paleosol: Mende F. 29 000 y. (1) N°2 Tata: 101 000 y. (2) Paleosol: Mende B.~120 000 y. (3)
0.4			Solymárium	Solymár	N°3	N°IIb N°III		Upper Old loess of Paks 2-3 paleosols	N°3 Buda, Kiscell: 190 000 y. (4) alluvial sand in old loess Paks: ~240 000 y. (3)
0.8		BIHARIUM	Mosbachium	Mosbach Betfia	N°4 ⊕	N°IV ⊕	↑	Lower part of old loess of Paks, 2 paleosols	N°4 Vértesszőlős: >350 000 y. (2) Paleosol: Paks PD_1, PD_2 both ⊕ ⊞
1.4			Cromerium	Cromer F.B.	N°5 ⊟⊞⊟	N°V	glacis formation of the mountains foreland	Lowermost part of old loess of Paks Pink colored sand	Oldest loess and paleosol (PDk) at Paks ⊖⊟⊖, at Dunaföldvár ⊕⊟
1.8		VILLÁNYIUM	Kislángium	Kisláng Villány-2.	N°6 ⊟⊖	N°VI	Lower lying foothill surface formation N°B	Red paleosol in N°6 Old alluvial fan of Kisláng	N°6 Dunaalmás Kisláng △
2.4			Beremendium	Beremend-5	N°7 ⊟⊞				N°7 Dunaalmás ⊟⊞
	PLIOCENE	BARÓTIUM	Csarnótanum	Csarnóta 2.	N°8 ⊟	N°VII	↑	Mottled clay; sand and red clay formation of Dunaföldvár: paleosol (Df_1 – Df_6) ↑	Upper Dunaföldvár Complex (Df_1-Df_4) ⊖⊕⊕ N°VII Kemeneshát; N°8 Dunaalmás ⊟
			Ruscinium str.	Serrat d'en Vacquer	N°9 ⊕	N°VIII	Climax of the pediment formation	↑ Correlative sediment of pedimentation of the Mátra foothill	N°VIII Kemeneshát gravel N°9 Köpite-hill ⊕ Pediment of Mátra Mts
5.4			Baltanum	Balta	N°10a ⊟⊞⊟	oldest alluvial fan of the Danube	foreland of Mátra, Villány and other mountains N°A	Optimum of the red clay formation, bentonite formation, sand formation ↓	Oldest red clays: Dunaföldvár ⊕⊖, Kulcs ⊕, Bag, Hatvan, Gyöngyösvisonta ⊞ N°10a Újhegy ⊟⊞⊟
	UPPER MIOCENE	BALTAVÁRIUM	Bérbaltavár	Baltavár-1			↑		sand formation of Gödöllő ⊖
			Hatvanium	Hatvan brickyard	N°10 △	in the foreland of mountains beginning of the formation of river system	Beginning of the pediment formation	↓ fluvio-lacustrine sand, delta, dune sand formation	N°10 Gerecse-Kőhegy, Várpalota Bérbaltavár sand ⊕△
			Sümegium	Sümeg-Kajmát					
10.0			Csákvárium	Csákvár cave	N°11 △	Marine terrace n°1			N°11 Széchenyi-hill △ n°1 Széchenyi-hill
		EPPELSHEIMIUM	Rhenohassium	Eppelsheim	N°12 △	Marine terrace n°2		delta gravel delta gravel	N°12 Szabadság-hill △ Travertine of Kapolcs n°2 Vértes-hill at Csákvár Szabadság-hill (Buda Mts)
			Bódvaium	Rudabánya-2					
			Monacium	München Flinz Sands		Marine terrace n°3		delta gravel and sand	n°3 Balaton - Upland (Balatonfüred) Buda Mts: Diósd, Kálla, Billege

Paleomagn. analysis made by:
◯ Pevzner, M. A.
☐ Márton, P.
△ Opdyke, N. D.

Th/U and ESR analysis made by:
(1) = Lab. Hannover, Moscow
(2) = Lab. Köln (Hennig et al.)
(3) = Lab. Debrecen (Borsy, Z. et al.)
(4) = Lab. Tallahassee/Florida (Osmond, J. K.)

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MICROSTRATIGRAPHICAL INVESTIGATIONS ON THE SHORE OF LAKE BALATON

S. Marosi, J. Szilárd

ABSTRACT

Latest results confirm authors' previous statements on the polygenetic origin of Lake Balaton basin through subsidence rhythmical in space and time during the Upper Pleistocene. This opinion has to be modified inasmuch as the gradual filling of the basin with water reached in Würm II the degree to produce shore sediments also in the eastern basin at 4 m above actual mean water level. This phase was followed by subsidence, regression, shore oscillation and the resulting soil formation, characteristic in the early Würm II-III interstadial. Then it was replaced by transgression, repeated soil formation, charcoal formation (dated 21,725±660 years B.P.) sedimentation related to oscillations and syngenetic semihydromorphous soil formation. Particularly from the the second half of Würm III strong regression, shore reshaping and slope deposition were typical. The latter reaches far into the present lake basin. They were accompanied by syngenetic solifluctional and cryoturbational processes. In the Holocene, the carving of the steep wave-cut margin, alluviation, bar formation and the intensification of human activities are the major factors controlling further geomporhic evolution.

INTRODUCTION

Lake Balaton is the largest and one of the most studied lakes in Central Europe.

L. LÓCZY Sen,in 1913 and J. CHOLNOKY in 1918, according to their theory of only two ice ages, stated that Lake Balaton originated in the Lower Pleistocene. In view of the polyglacial theory B. BULLA and A. KÉZ in 1943 concluded that the lake was formed in the Riss-Würm interglacial. In the 1950's B. ZÓLYOMI (1952) considered the formation of the lake-basin to have occured during Würm III, basing his supposition on pollen analytical studies, while J. SÜMEGHY (1955) assumed that the lake existed only from the Early Holocene onwards after he examined the available stratigraphic evidence. The authors of this paper began their investigations in the area in the middle fifties. We have succeeded to resolve these former conflicting views by suggesting a polygenetic theory of lake formation, a rhythmic development both in space and time (MAROSI, SZILÁRD 1958, 1977, 1981). Evidence found to confirm this theory is dealt with in several papers (SZILÁRD 1965a, 1965b, 1967; MAROSI 1960, 1965, 1969, 1970).

Meanwhile, at the Paris conference of the INQUA Commission in 1969 J. SZILÁRD presented a brief summary about the origin and evolution of the lake basin (SZILÁRD 1970).

The palynological evaluation of lake-bottom sediments (ZÓLYOMI 1980) and stratigraphical analysis of coastal sequences were conducted by different methods and approach, yet the findings conform desirably.

Our microstratigraphic and chronological studies around the lake gain wider perspective if we add that most of Hungary's topographic features, namely the plains and hilly landscapes are composed of thick deposits of basin sediments. Above the more compact Pliocene sea and inland-lake sediments, there lie Central Europe's most widespread Pleistocene loess and sand sequence alluvial fans and terrace formations, blown-sand areas etc. Quaternary research has more than a century old tradition in Hungary. Researchers' attentionn was, and is frequently focused on examining the chronological and other geomorphological aspects of these sedimentary environments. Chronological evaluation of these sediments and detailed analyses and syntheses by modern methods yielded significant results in the past decades. Consequently, a more precise chronological classification became possible, and simultaneously attempts were made to correlate sediments and paleosols subdividing them both locally, regionally and on an international scale (RÓNAI, 1969,KRETZOI, KROLOPP 1972; PÉCSI 1969, 1970, 1973, 1975 etc).

Among the most important sedimentary profiles in Hungary, the coastal sequences, the subject of this paper, offer a unique opportunity to study the special ecological conditions under which they were formed. The local circumstances are reflected in the complex genesis of these sequences.

Our detailed microstratigraphical investigation on the Lake Balaton shore revealed the stages in the evolution of the lake-basin. It enabled us to determine the period in which the basin was formed. These sediments served as the basis for a detailed chronological subdivision of the Late Pleistocene. If correlated with the appropriate sediment sequences of those continental sediments that were formed under different ecological conditions, the ecological conditions, the ecological anomalies related to lacustrine sedimentation, wave action and the oscillation of the water level, become obvious.

Relief Development Prior to the Formation of the Lake-Basin

During the transitional Plio-Pleistocene period fluvial erosion and accumulation came to predominate over the former deposition of inland-lake sediments. The rivers often changing their courses, wandered over large areas. This process characterized fluvial activity until the Early Pleistocene subsidence of the Balaton basin. Originally, rivers were flowing across Western Hungary from the Alpine Carpathian catchment basin towards the Croatian-Slavonian depression (SÜMEGHY 1955; Fig. 1). During the Lower and Middle Pleistocene due to the gradual emergence of the Transdanubian Mountains these watercourses lost their catchment areas lying north of the Mountains. Their base level of erosion also shifted to the younger upper Kapos-Kalocsa graben-like depression situated 50 km south of the present-day Lake Balaton (MAROSI 1960; Fig. 2, 3). Fluvial and correlative sediments from the Mountains testify the existence of this former fluvial system (Fig. 3). In the graben at some places these sediments are more than 100 m thick and are also there in the N-S valleys of today's hilly region south of the lake. Relict forms of these former fluvial and correlative mountain sediments are the terrace systems of the north-south valleys (SZILÁRD 1965a; Fig. 4). A marked characteristic of the terrace system is that, while an

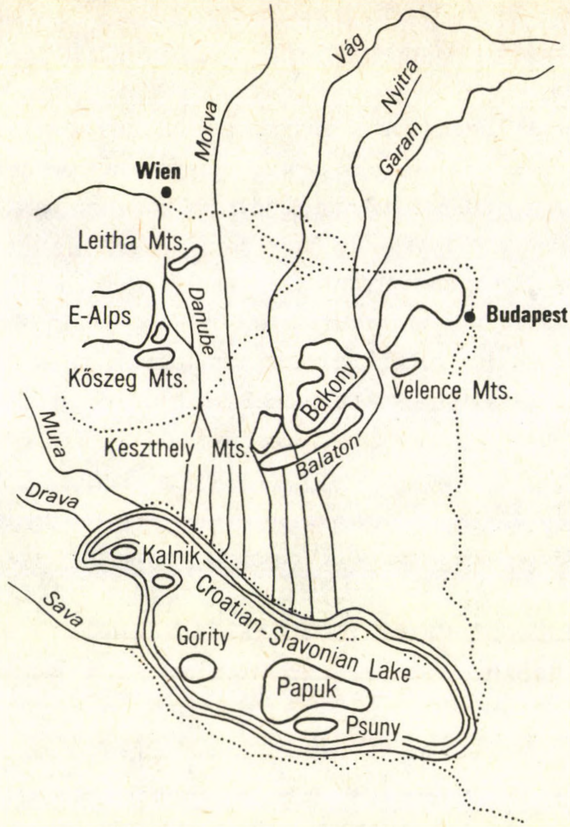


Fig. 1: Hypothetical map of drainage in Transdanubia during the Piacensian Stage (according to SÜMEGHY 1955)

older and higher terrace slopes definitely to the south, two lower terraces 'break off' at mid-valley, both incline from a mid-valley watershed to the north on one side and to the south on the other. The northern sections of these lower terraces therefore are orientated towards the new and younger base level of erosion of the Balaton basin. This later subsidence of the basin radically modified the whole of the pre-existing drainage.

In the upper Kapos-Kalocsa graben-like depression C o e l o d o n t a a n t i q u i t a t i s fossils were found in the upper layers of the fluvially transported correlative sediments from the Uplands (SZILÁRD 1967). These provide important evidence about the thorough reshaping of relief and about the origin of the Balaton basin. Immediately over the correlative fluvial

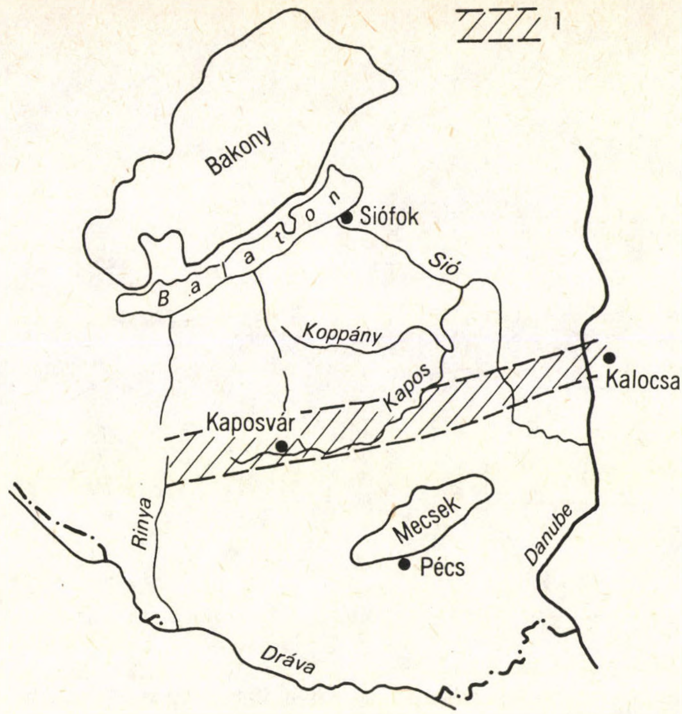


Fig. 2: Geomorphological sketch of the environs of Lake Balaton.
1 = Upper Kapos-Kalocsa depression

sediments there are Würm loess series with fossil soils subdividing them. This proves beyond doubt that the fluvial system with rivers flowing across open spaces to the south could only have existed up to the beginning of the Upper Pleistocene, when it was halted by the subsidence of the lake-basin. After this time only eolian sedimentation occurred in its former base level of erosion, in the upper Kapos-Kalocsa depression. According to this turn of events, the Balaton basin resembling the present-day lake cannot be older than Würm period is available from lake-shore sediment profiles and can be proved by microstratigraphic analysis.

Microstratigraphical Investigation of the Sediment Profile near Siófok

Among other sediment profiles examined in great detail, the one on the southern shore of Lake Balaton near Siófok is of special importance in

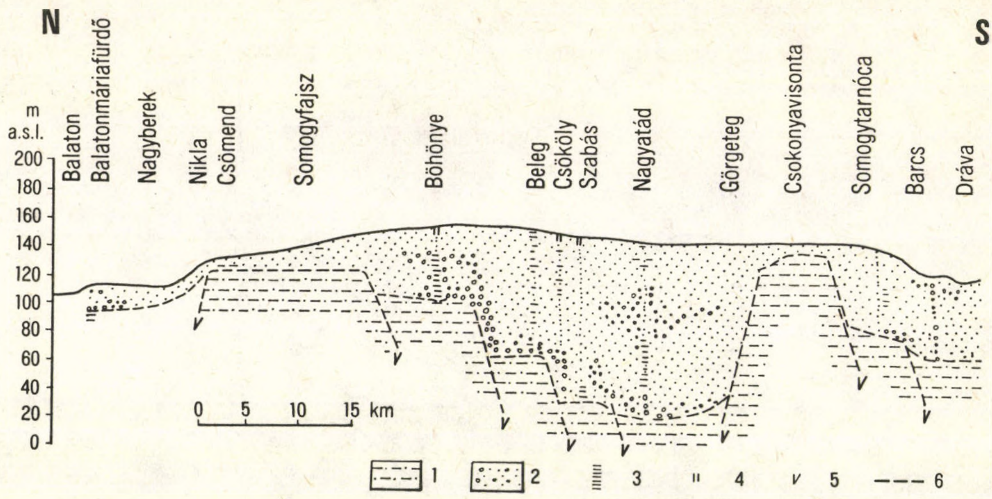


Fig. 3: A north-south cross-section from Lake Balaton to the Dráva river (constructed by MAROSI).

1 = Pannonian sandy, clayey, silty sedimentary layers the upper part of which consist of Upper Pliocene cross-bedded sands of varying thickness; 2 = Pleistocene clayey, silty, sandy and detrital fluviatile and lake sediments with wind-blown sand cover and loess patches; 3 = clay from core sample; 4 = loess, sandy loess from core sample; 5 = probable fault zone; 6 = Plio-Pleistocene stratigraphical boundary

providing clues for lake formation. Vertically, the exposure extends form 104,1 m (1.5 m below the present mid-water level of the lake) to 121 m. (Pict. 1) The site of the three profiles is illustrated in detail in Fig. 5. Pict. 2 illustrates the gently arching sedimentary layers of an infilled dell, and Pict. 3 a section of this infilled dell showing minutely interlayered loess, sand, and gravel beds slightly dislocated as a result of compaction. Pict. 4 depicts the upper section, a stratified loess series interlayered with sands and gravels, the holes were made by sand martins. Sections of the exposure (profiles I-III) can be arranged in a stratigraphic order. The oldest layers are the pre-Balaton fluvial deposits overlying the

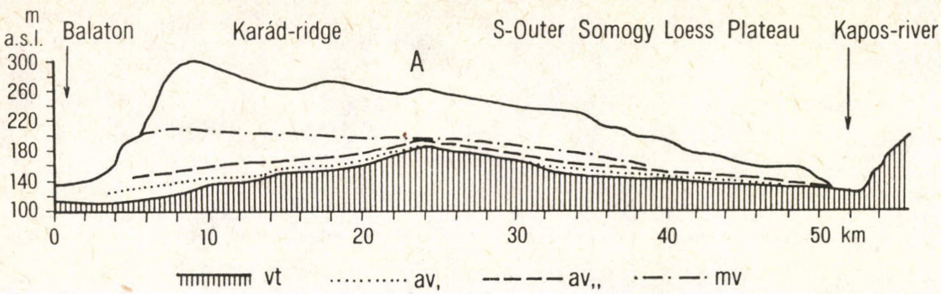
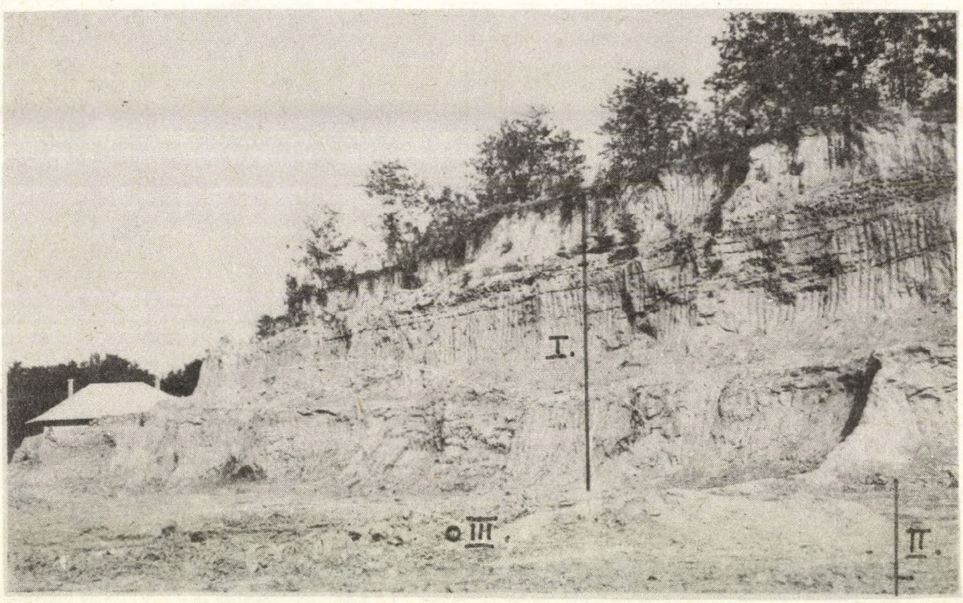


Fig. 4: Longitudinal section of the Somogytúr-Orci meridional valley (constructed by SZILÁRD). A = valley watershed, vt = valley bottom, av, = lower level of valley side slope, av,, = middle level of valley side slope, mv = shoulder of valley side slope



Pict. 1: General outlook of the Balatonszabadi-Sóstó exposure, detailed profiles marked I-III are shown on Figure 5.

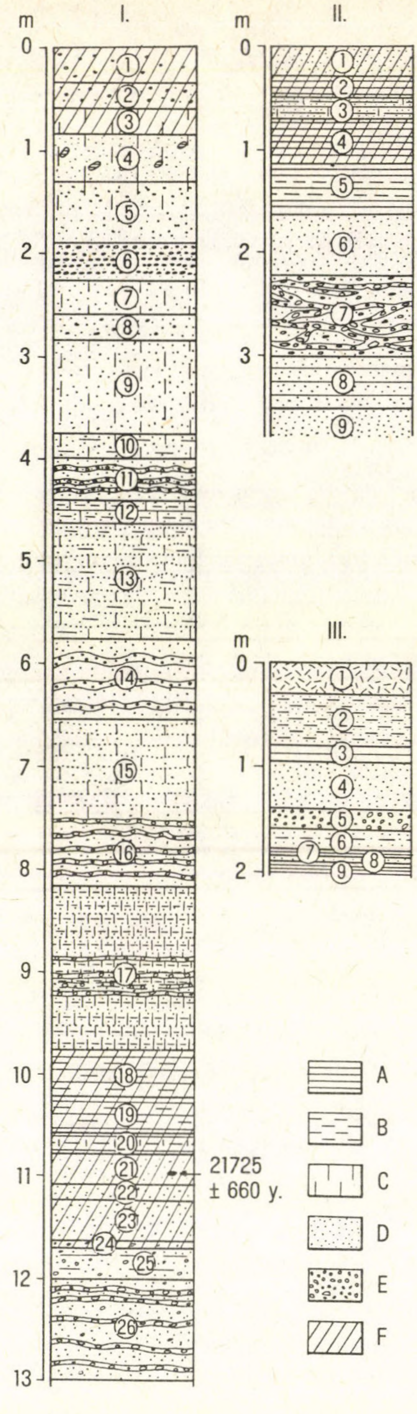
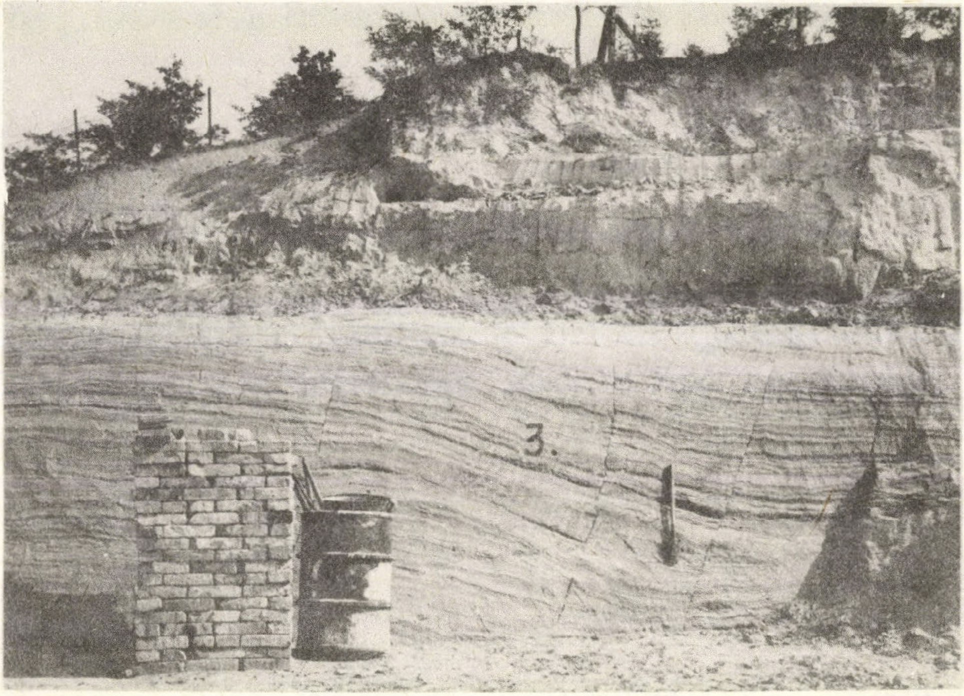

I.
II.
III.
m
21725
± 660 y.
A
B
C
D
E
F

Fig. 5: Profiles I-III of the Balatonszabadi-Sóstó exposure. Profile I: 1-4 = chernozem soil, 5-17 = slope sediment complex, 18-24 = fossil soil complex, 25-26 = sandy, detrital complex. Profiles II: 1-4 = fossil soil complex, 5-9 = Balaton Lake sediments. Profile III: 1 = recent rain wash deposit, 2-3 = Balaton Lake sediments partly of Pannonian and fluviatile origin, 6-9 = Pliocene (Pannonian) sediment, A = clay, B = silt, C = loess fraction, D = sand, E = fine gravel, detritus, F = soil



Pict. 2: Part of the north-eastern section of the exposure; gently arching sedimentary layers of an infilled dell

Pannonian sediment sequences. These are represented as the 2-5 layers of profile III and the 8-9 layers of profile II. shown in Fig. 5. The 2-7 beds of profile II are a lacustrine Balaton sequence rich in broken shells of

Pict. 3: A section of the infilled dell (see Pict. 2) showing interlayered loess, sand and gravels beds

snails and molluscs. The cross-section at right angles to the present coastline reveals a specific arching of these layers. Overlying these bar-bay formations there is a hydromorphous and semihydromorphous double soil complex indicating a regressional period. It is also represented by the 4th and 3rd layers of profile II. Semihydromorphous soils developed on the shore as a result of the lowering of the water level accompanied by slight deflational activity and dell formation. This is shown by the layers 21-26 in profile I. The subsequent occurrence of forest soil formation indicates an increase in the amount of precipitation and moisture. Eventually, the lake inundated these forested areas and destroyed the trees. This event ensured the supply for the intensive accumulation of charcoal in layer 21. It is

Pict. 4:Part of profile II showing the level in which fossil soils were formed

21725$\pm$660 years old according to the C^{14} method and is situated 9,8 meters below the top of the exposure, and 7 m above the presen mid-water level of the lake (high water level did not exceed 4-5 m during the Holocene).

The layers 1-17 situated above the soil complex are slope deposits of interlayered sandy, marshy and loess strata whith gravel beds of variously sized pebbles. These deposits are partly the product of the dell system orientated towards the lake. The finely cryoturbated and specially stratified layers were formed during the Pleistocene as a great number of small icewedges also testify. The presence directly above these, of pseudomicaelian chernozem soils signal that we cannot account for postglacial sedimentation in this exposure. A yearly accumulation of 1 mm can be supposed beginning with the stratum dated by the radiocarbon method.

According to our interpretation and available data, the significant 21st layer with enclosed charcoal remnants is a later humified hydromorphous A horizon of the former B horizon of the semihydromorphous forest soil found immediately below. The formation of both of these soils is dependent on a particulat water level and morphoclimatic condition, though they are the product of two different stages. Thus, it is reasonable to suppose for the purpose of a chronological classification that starting with the lower lacustrine sediments, there is no significant stratigraphic hiatus in the sequence. The following stages of development can be distinguished in the evolution of the lake:

1. Lying discordantly over the Pliocene sedimentary substratum only thin beds of fluvial gravelly sands are preserved from the pre-Balaton era, the first and longer period of the Quaternary.

2. The fluviatile sediments found today along the shore and in long stretches along the valleys running in a meridional direction, or at some places forming deluvial slopes, were already present when the subsidence of the Balaton trench began. Subsidence was accompanied by a radical alteration of drainage, rivers started flowing towards the basin, and numerous river captures took place. There is no indication of this process in the exposure we have examined at Balatonszabadi-Sóstó, since it lies off the course of the former N-S valleys.

3. The continuously subsiding Balaton basin was gradually filled with water, so that lacustrine sediments can be traced from this period (the first important transgression occurred during Würm III).

4. Regression connected with the strong subsidence of the basin continued still at a relatively high water level, and hydromorphous and/or semihydromorphous soil formation went on along the nearby shores where it was accompanied by a slight oscillation of the water level. As regression advanced further, the relative relief between the base level of erosion and shores began to increase, erosional processes channelled towards the lake became more active, especially linear dell formation. The afforestation of the area resulted in semihydromorphous forest soil formation (the first part of the Würm II, Würm III interstadial).

5. During the rest of the Würm III interstadial the subsidence of the basin slowed down, and an increase in the amount of precipitation resulted in a new transgressional period (The profile shows traces of flooding in this period.) The forest was destroyed, hydromorphous soil formation took place once more, and charcoal was produced (21725±660 years). Although the

transgressional trend was dominant, the oscillation of the water level was frequent and thus the accumulation of sediments accompanied by a syngenetical semihydromorphous soil formation.

6. During the second half of Würm III the emergence of the coastline was due to marked regression. Periglacial processes (pluviation, solifluction, cryoturbation) became active, and there was a rapid accumulation of mixed sediments.

7. From the periglacial period onwards under the more humid but constantly changing climatic conditions the oscillating waters destroyed the former shove slopes, and a steep wave-cut rim was formed along the coast. Valley incision by rivers running towards the lake is evident in this period, and dells take a different course than formerly. The rhythmic climatic changes and tectonic movements of the Holocene did not affect the later development of the shoreline significantly. Their overall effect was the formation of a gently sloping alluvial plain and a system of low bars situated between the high banks and the water margin; along the axis of the small derasional valleys the channels exhibit alternating stages of erosion and deposition.

8. Holocene development characterized by climatic and morphosturctural diversity cannot be detected and differenciated on the basis of any one profile examined. However, it is possible to draw some conclusions from the study of the evolution of bar systems, alluvial plains and sediments in the former lagoons today marshy groves.

'Pseudomicaelian' chernozem soils developed to the south of the wave-cut rim, with a natural vegetation of steppes on sand-plains and meadows on loesses.

9. The most recent relief development is characterized by human activity. The environs are increasingly used for agricultural purposes with a resultant destruction of soils and slopes. The shore has been built up, trees are planted, and there is a growing need for the protection of the shorelines.

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DISTRIBUTION OF SOME PLEISTOCENE MOLLUSC SPECIES IN HUNGARY

E. Krolopp

ABSTRACT

The study of the regional distribution of Pleistocene Molluscs substantially contributes to knowledge on Quaternary stratigraphy and paleoenvironments. Author presents the preliminary results of his investigations of the mollusc species *Corbicula fluminalis* /MÜLL./, *Helicigona* banatica /ROSEM./, *Granaria frumentum* /DRAP./ and *Cochlodina laminata* /MONT./.

INTRODUCTION

A zoogeographical difference was recognized by JÁNOSSY, D. /1979/ between the Upper Pliocene - Lower Pleistocene vertebrate fauna of the North Hungarian Mountains and of the Villány Mountains, South Hungary. This recognition encouraged me to start the study of regional aspects of Pleistocene molluscs. Though my investigations of this type are at the very beginning, some valuable data have been obtained which are presented below.

DISCUSSION

The bivalve *Corbicula fluminalis* /MÜLL./ is known from Late Lower Pleistocene sediments in the southern and southeastern parts of Hungary /KROLOPP, 1978b/. The few occurrences in other regions are of different ages. It was found south of Lake Balaton, at Szabadhidvég in the older sediments /Upper Villányium KROLOPP, 1978a/, at Eger in younger layers /Riss-Würm interglacial; see HABLY, et al. in prep./ while the exact age of the occurrence at Zánka /northern shore of Lake Balaton/ is unknown. Based on the published data and on the enclosed sketch map /Fig. 1/ it can be stated that in the Carpathian basin the area of this species of stratigraphic significance extended northwards and westwards to the southern-southeastern corner of Hungary in the Lower Pleistocene /Lower Biharium/. The limit of the area was probably controlled by climatic factors.

Among the Upper Pleistocene examples, the gastropod species Helicigona banatica /ROSEM./ is to be mentioned first. The distribution of this index fossil of the Riss-Würm interglacial /LOZEK, 1964/ is restricted to the northern-northeastern margin of Hungary /Fig. 2/, it is absent from the Transdanubian sediments of the same age. The lack of the species in Transdanubia can be explained by the more continental climate of the region compared to the Norh-Hungarian Mountains /and also to Northwest Europe-KROLOPP, 1969/.

The gastropod species Granaria frumentum /DRAP./ is widespread in the older Pleistocene formations of Hungary. Neverthelles, in the Upper Pleistocene, i.e. Würm loesses of Hungary it occurs only in the southern part of the country /Fig. 3/. The species is thermophilous and xerotherm and these features are responsible for extension of its area outlined above. Some occurences differing from this pattern indicate the formation in question may be older than Würm.

The species Cochlodina laminata /MONT./ lives in woody-shrubby areas. in Upper Pleistocene /Würm/ sediments it is known only from the southern part of Transdanubia /Fig. 4/. The conclusion can be drawn that in the Würm this area was more densely vegetated than other lowlands and hills of the country. The woody-shrubby regions could develop under favourable precipitation conditions. Nevertheless, the possibility cannot be excluded that the Danube that run roughly along its recent course in the younger phases of the Upper Pleistocene provided ecological conditions for vegetation similar to the recent gallery forests together with the related gastropod species.

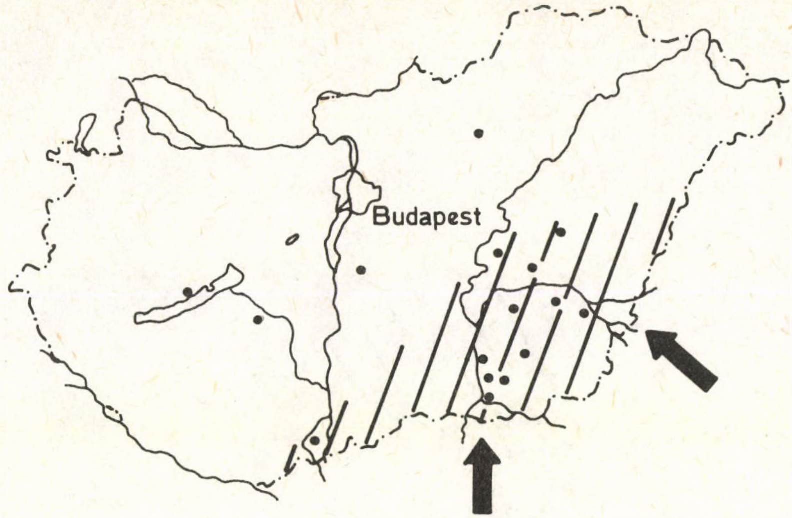

Fig. 1: Localities and Late Lower Pleistocene distribution of C o r b i c u - l a f l u m i n a l i s /MÜLL./ in Hungary

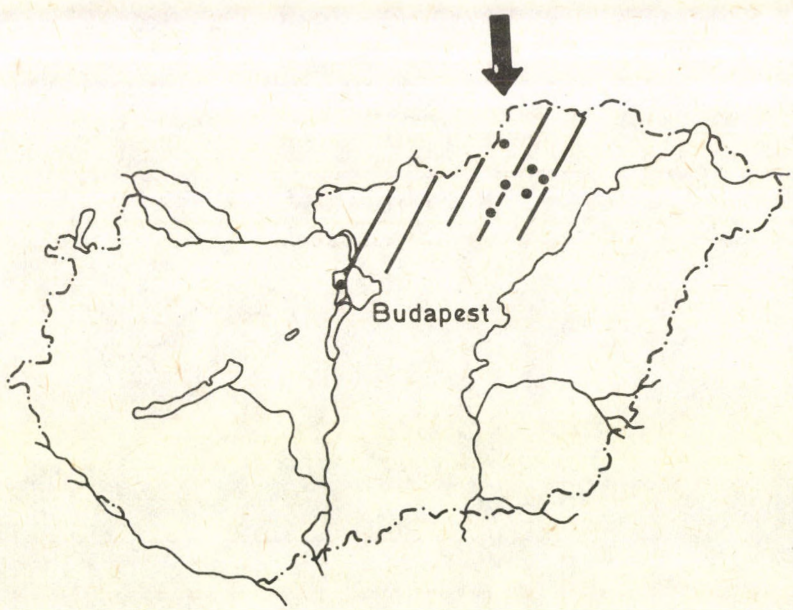

Fig. 2: Localities of H e l i c i g o n a b a n a t i c a /ROSSM./ in Hungary in the Pleistocene, Riss-Würm interglacial

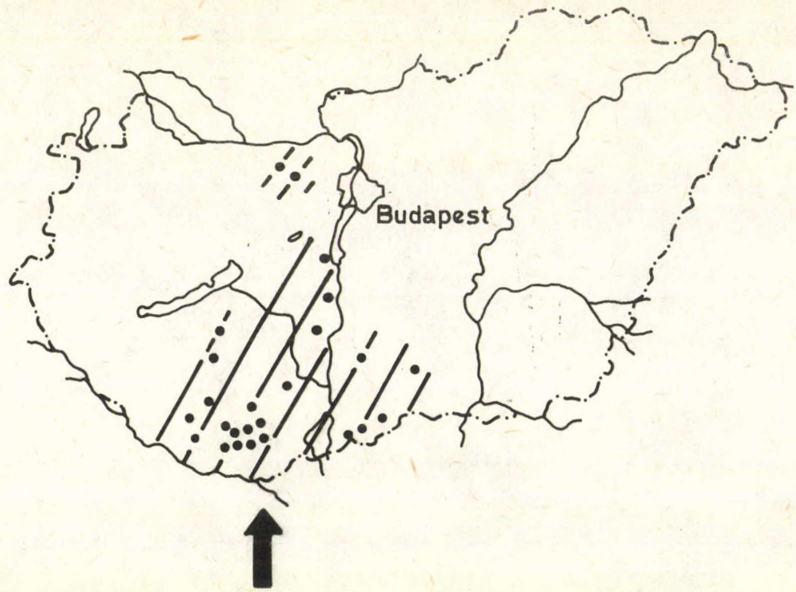

Fig. 3: Würm localities of G a n a r i a f r u m e n t u m /DRAP./ in Hungary

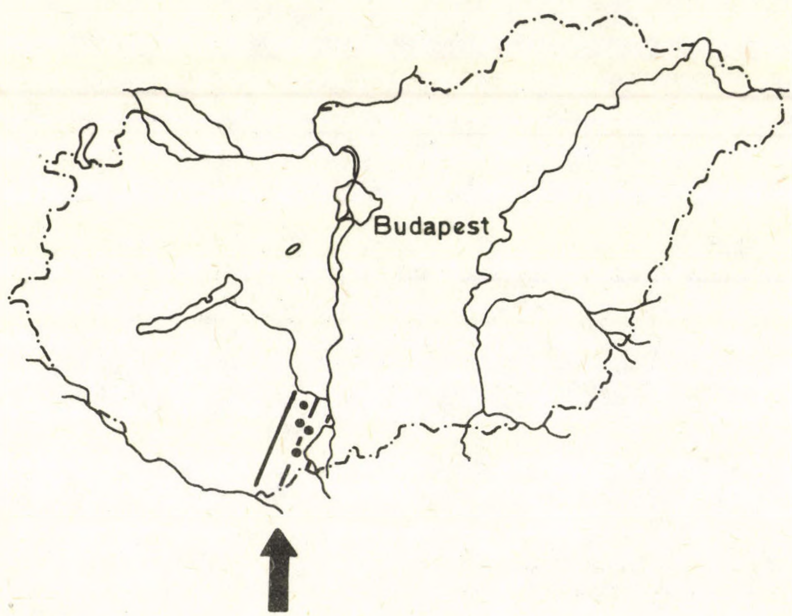

Fig. 4: Occurrence of C o c h l o d i n a l a m i n a t a /MONT./ in Hungarian Würm loesses

CONCLUSION

Based on these facts and on examples not discussed here it can be claimed that the zoogeographical characteristics reflected in the distribution of Pleistocene mollusc species within the Carpathian Basin can be and should be taken into account in

1. paleoecological and paleogeographical reconstructions;
2. stratigraphy.

In possession of exact knowledge on the areal distribution of species of stratigraphic significance misinterpretation in stratigraphic evaluation based on the presence or absence of certain species can be avoided. However, to fix exactly the regional features of the Pleistocene distribution of certain species caused by ecological factors, the detailed analysis of occurrence data is needed.

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Budapest, 1988

POSSIBLE CORRELATIONS BETWEEN LOESS AND CAVE DEPOSIT STRATIGRAPHIES FOR THE UPPER PLEISTOCENE IN HUNGARY

Á. Ringer

ABSTRACT

Based on lithological, paleontological, paleobotanical, archaeological, and radiometric evidence, the paper attempts to correlate chronologically the loess and cave deposit stratigraphies for the Upper Pleistocene of Hungary and also demonstrates some major implications of this complex chronostratigraphy for European correlation.

INTRODUCTION

As in general, also in the middle Danubian basin, Upper Pleistocene subaerial and cave deposits are widest-spread, best-developed and most readily subdividable. In our young loess series the last interglacial soil is only overlain by interstadial paleosol (PÉCSI, 1985; PÉCSI et al., 1977). Their stratigraphical position, expression, paleopedological properties, and absolute age parallel with the same formations in other parts of Europe (PÉCSI, 1978). From syngenetic subaerial and cave deposits, vertebrate paleontological (KRETZOI, 1968; KRETZOI-VÉRTES, 1965; KRETZOI-PÉCSI, 1982; JÁNOSSY, 1979; KORDOS, 1986; VÖRÖS, 1984, 1987) malacological (KROLOPP, 1982), and paleoanthropological (THOMA, 1968) data and Upper Pleistocene flora remnants (ZÓLYOMI, 1953; STIEBER, 1957; JÁRAI-KOMLÓDI, 1968; SKOFLEK, 1980; PASHKEVICH, 1979; URBAN, 1980, 1984) and Middle and Upper Paleolithic archaeological finds (GÁBORI-CSÁNK, 1968, 1970, 1983; GÁBORI, 1964, 1976, 1984; VÉRTES, 1965; T. DOBOSI, 1975; T. DOBOSI et al. 1983; RINGER, 1983, 1986) were recovered.

The results of absolute dating of young loess and cave deposits fix their formation at 125,0±20 ka and 11,3±0,5ka BP (BORSY et al. 1979; BUTRYM-MARUSZCZAK 1984; PÉCSI, 1975, 1985; GEYH et al. 1969; GÁBORI-CSÁNK, 1970; KROLOPP, 1977; VÉRTES, 1965).

In the paper the litho- and biostratigraphical divisions of Upper Pleistocene subaerial and cave deposits in the Middle Danubian basin are presented. An attempt is made at chronostratigraphical correspondence and some possible international correlations and important paleogeographical-paleoecological implications are also mentioned.

A CHRONOSTRATIGRAPHICAL SYSTEM OF YOUNG LOESS AND CAVE DEPOSITS

Lithostratigraphy

The most important loess profiles in Hungary are described from the Great Hungarian Plain and marginal hill regions of the Hungarian Mountains at 100-280 m above sea level. The major explored caves open in the karst mountains at 150-750 m altitude (Fig. 1).

Loess

The Upper Pleistocene loess of the Middle Danubian basin is divided into the Mende-Basaharc series and the Dunaújváros-Tápiósüly series (PÉCSI, 1965a, 1965b, 1985; HAHN, 1977).

The Mende-Basaharc loess series is the lower member of young loess, it is of last interglacial, early and middle glacial age.

The loess series of 20-25 m thickness described from the wall of the former Mende and Basaharc brickyards (PÉCSI, 1965a, 1965b) contains last interglacial formations with the overlying loess horizons l_5, l_4 and l_3 with the paleosols BA, BD_2, BD_1, MF_2 and MF_1 (Figs. 2-3). The loess series is also divided by soil sediments, dell fills and periglacial features. Dust

Fig. 1: The most important loess profiles and major explored caves in Hungary. – a = major loess profiles, b = major explored caves: 1 = Subalyuk; 2 = Lambrecht; 3 = Büdöspest; 4 = Istállóskő; 5 = Szeleta; 6 = Peskő; 7 = Balla; 8 = Puskaporos niche; 9 = Pongor-lyuk; 10 = Pilisszántó niche I; 11 = Kiskevély; 12 = Bivak; 13 = Remete Upper.

accumulation took place, in our present knowledge, for ca 100 ka.

The upper member of young loess, the Dunaújváros-Tápiósüly series is of late glacial age.

The 5-10 m thick series (PÉCSI, 1975; HAHN, 1977) has two loess horizons (l_2 and l_1), which mostly developed under cold arid climate over a relatively short time span (15-17 ka). Only poorly developed, embrionic soils (H_2 and H_1) as well as small dell fills and periglacial features (Fig. 2-3).

Cave deposits

Litho- and biostratigraphic data of 13 layers from ca 60 caves in Hungary explored since 1906 have been analyzed. The lithological evidence,

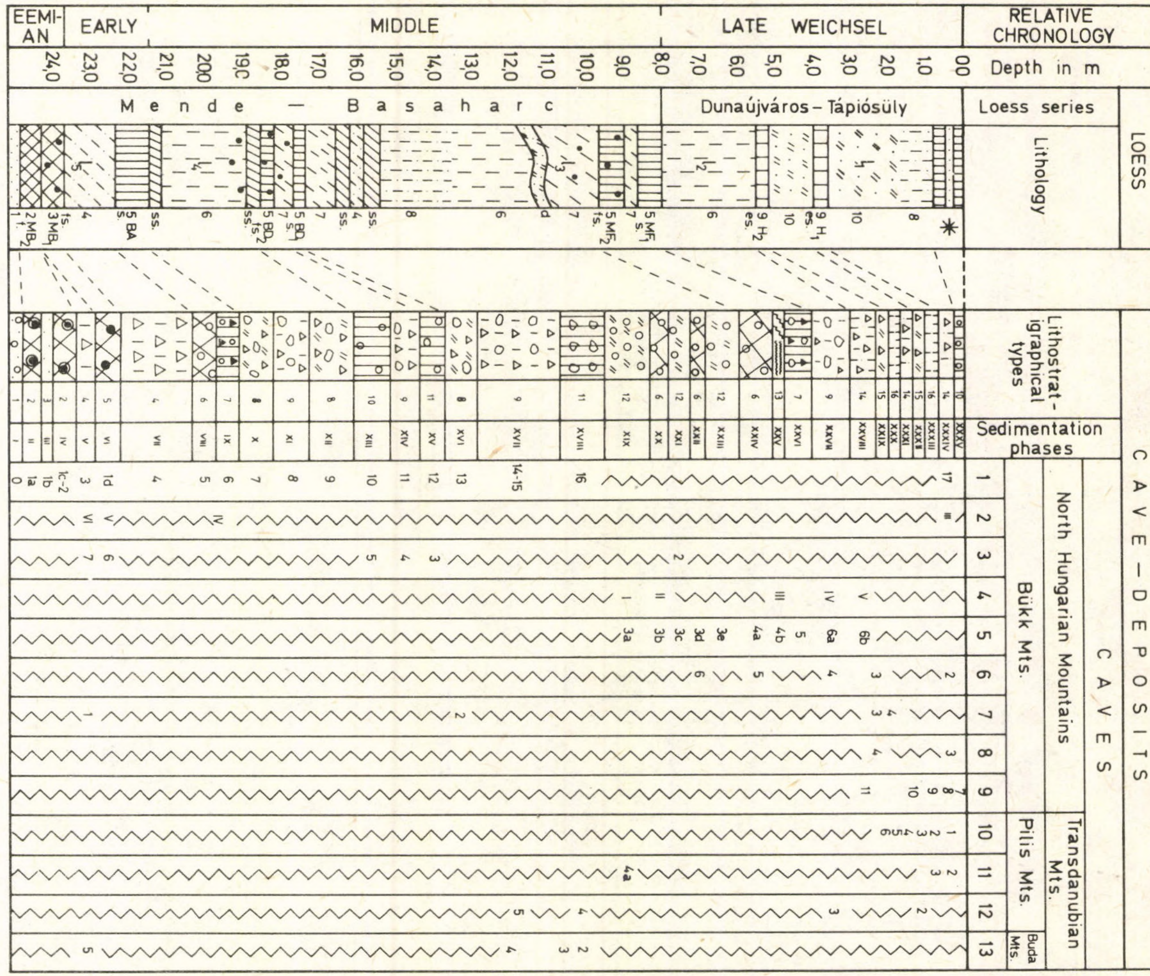

EEMI-AN
EARLY
MIDDLE
LATE WEICHSEL
RELATIVE CHRONOLOGY
Depth in m
Mende — Basaharc
Dunaújváros - Tápiósüly
Loess series
Lithology
LOESS
Lithostratigraphical types
Sedimentation phases
CAVE-DEPOSITS
CAVES
North Hungarian Mountains
Bükk Mts.
Transdanubian Mts.
Pilis Mts.
Buda Mts.

Fig. 2: Correlation between the general profile of Upper Pleistocene loess in Hungary and the corresponding sequence of major caves in Hungary.

Loess

Lithology: 1 = fluvial sand; 2 = Mende Base paleosol, lower part; 3 = Mende Base paleosol, upper part; 4 = sandy slope loess; 5 = interstadial paleosol; 6 = stratified loess; 7 = stratified slope loess; 8 = sandy loess; 9 = embrionic soils; 10 = stratified sandy slope loess.

l_1, l_2, l_3, l_4 and l_5 = loess packets, d = dell.

Paleosol types: f = lessivated brown forest soil with submediterranean climatic influence (MB_2), fs. = forest steppe soil (MB_1, BD_2, MF_2), s. = chernozem like steppe soil (BA, BD_1 and MF_1), ss = Soil sediment, es. = embrionic soils, * = the late glacial Bölling and Alleröd interstadial soils in the Nyirség test area identified by Z. BORSY.

Cave Deposits

For numbers of caves see Fig. 1. Other numbers show the sediment sequences of caves.

Lithostratigraphycal types: 1 = Cave stream sand and gravel; 2 = decalcified cave soil or soil sediment of high clay content containing few, highly corroded limestone debris and Fe precipitations, sub mediterranean climatic influence (Geochemistry: SiO_2 - 66.46 %, Al_2O_3 - 19.28 %, Fe_2O_3 - 6.52 %, TiO_2 - 0.40 %, $CaCO_3$ - 0.82 %); 3 = cave stream; 4 = cave loess with few and small frost shattered limestone debris; 5 = cave soil or soil sediment of high clay content containing limestone debris of mostly corroded surface and sharp edges (Geochemistry: SiO_2 - 52.92 %, Al_2O_3 - 15.30 %, Fe_2O_3 - 6.67 %, $CaCO_3$ - 19.00 %); 6 = cave soil or soil sediment of relatively high clay content and with few and small frost shattered limestone debris (Geochemistry: SiO_2 - 42.18 %, Al_2O_3 - 24.50 %, Fe_2O_3 - 5.65 %, TiO_2 - 0.35 %, $CaCO_3$ - 11.61 %); 7 = cave soil or soil sediment of relatively high clay content and with limestone debris; particle have rounded edges and corners (Geochemistry: SiO_2-

56.37 %, Al_2O_3 - 14.86 %, Fe_2O_3 - 5.10 %, $CaCO_3$ - 10.00 %); 8 = cave loess enriched in allochtonous fine fraction and coarse limestone debris; frost shattered debris is partly rounded (Geochemisty: SiO_2- 57.50 %, Al_2O_3 - 12.10 %, Fe_2O_3 - 4.58 %, $CaCO_3$ - 27.00 %); 9 = layers of fine and coarse limestone debris embedded in cave loess (Geochemistry: SiO_2 - 42.44 %, Al_2O_3 - 20.66 %, Fe_2O_3 - 6.64 %, TiO_2 - 0.22 %, $CaCO_3$ - 13.71 %), 10 = black, rendzina-like cave soil or soil sediment with few and small limestone debris (Geochemistry: SiO_2 - 47.40 %, Al_2O_3 - 13.97 %, Fe_2O_3 - 4.97 %, $CaCO_3$ - 31.00 %); 11 = weakly developed, black rendzina-like cave soil or soil sediment with much limestone debris mostly of large size (Geochemistry SiO_2 - 30.60 %, Al_2O_3 - 7.84 %, Fe_2O_3 - 2.92 %, $CaCO_3$ - 57.00 %), 12 = cave loess with mostly rounded frost shattered limestone debris (Geochemisty: SiO_2 - 22.70 %, Al_2O_3 = 12.14 %, Fe_2O_3 - 3.09 %, $CaCO_3$ - 58.00 %); 13 = travertine precipitation, hiatus; 14 = sandy cave loess with angular limestone debris; few large blocks among limestone debris (Geochemistry: SiO_2 - 47.30 %, Al_2O_3 - 9.19 %, Fe_2O_3 - 3.55 %, $CaCO_3$ - 30.00 %); 15 = the same gelisolifluctional loess but with more limestone debris, 16 = subarctic cave soil or soil sediment (Geochemistry: SiO_2 - 46.50 %, Al_2O_3 - 9.80 %, Fe_2O_3 - 5.20 %, $CaCO_3$ - 34.00 %)

absolute dating, paleontological and archaeological data allowed to identify 35 layers, i.e. paleoclimatological-paleoecological phases in the succession most probably embracing the whole Upper Pleistocene (Figs. 2-3).

The loess series Mende-Basaharc and Dunaújváros-Tápiósüly best correlated with the fills of the caves Subalyuk, Lambrecht, Büdöspest, Istállóskő, Szeleta in the Bükk Mountains and with the Pilisszántó niche in the Pilis Mountains. Less complete fills only correlate in some sections (Fig. 2).

In the caves periglacial limestone debris layers with cave loess and loess loam and interstadial, mostly rendzina-like cave soils or soil sediment. In both sediment sequences paleoclimatic (erosional, gelisolifluctional, cryoturbational-cryodeformational) phenomena are characteristic and paleoecological (fauna and flora) features of the same climatic type can be correlated and archaeological cultures appear in similar stratigraphic positions (Fig. 3).

Lithostratigraphic correlation between loess and cave deposits

Upper Pleistocene loess (20-25 m) and cave deposits (19-26 m) have similar thicknesses and both can be divided into two parts lithologically– paleopedologically (Figs. 2-3).

The Mende-Basaharc loess series and the cave sedimentation phases I-XXVI represent the last interglacial, lower and middle glacial, while the Dunaújváros-Tápiósüly loess series and the cave sedimentation phases XXVII-XXXV can be placed to the late glacial (Figs. 2-3).

The interglacial complex, present at the base of both series, consists of an erosion period and a brownish red or red forest soil (MB_2) or its cav soil and soil sediment counterpart (sedimentation phases II and IV).

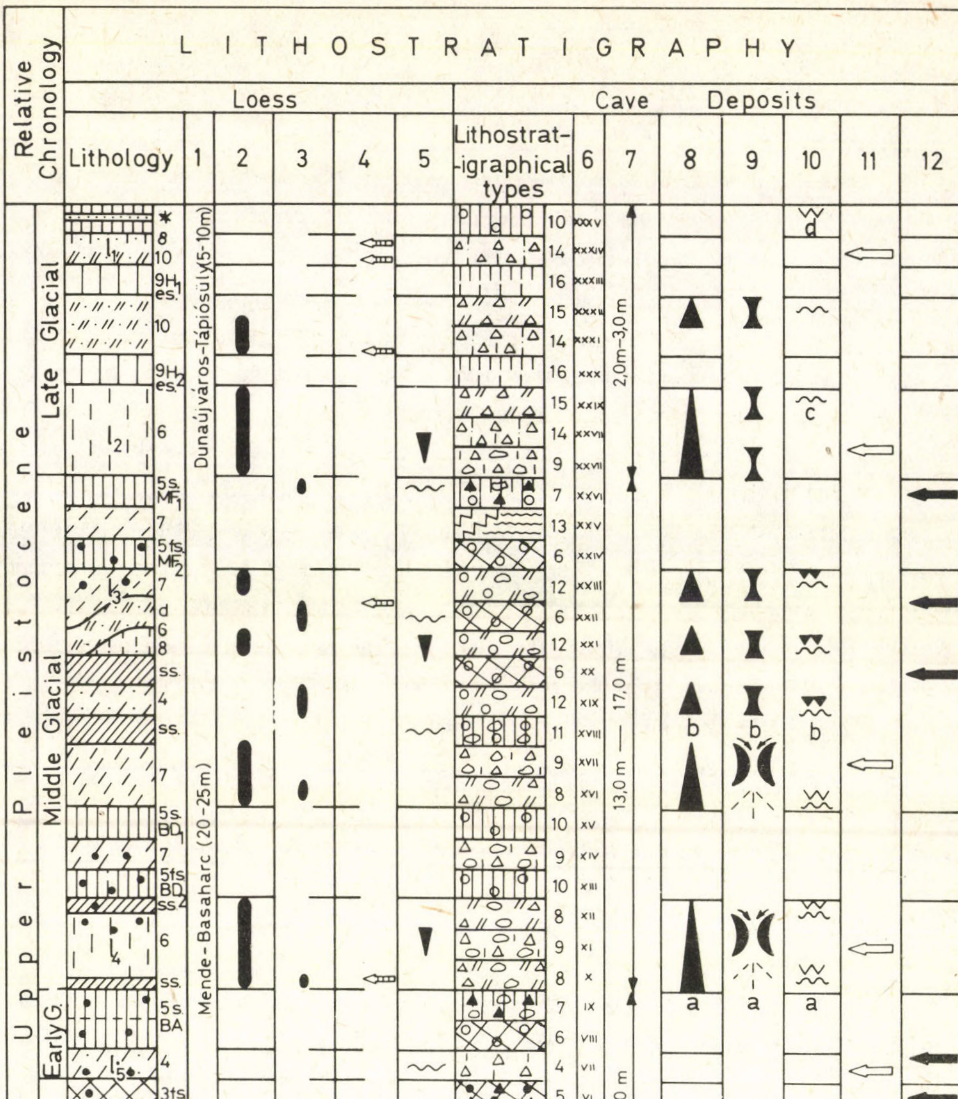
For the early glacial two stadials can be identified in both sequences. In loess the thin loess substrate of the forest steppe soil MB_1 formed during the first stadial, the loess horizon l_5 and in caves the periglacial cave sediment of sedimentation phases VIII and X-XII in the second.

Between the first and second stadials the upper member of the MB soil complex developed, while after the second stadial the well-developed fossil chernozem BA came into being. They may correlate with the cave soils or soil sediments of sedimentation phases VIII-IX. During the stadial at the beginning of the middle glacial loess horizon l_4 and the deposits of cave sedimentation phases X-XII developed.

The gelifluctional soil sediment at the bottom and top of loess l_4 may date back to the same cold and humid climatic oscillations when soliflucional cave loess formed during the cave sedimentation phases X and XII (Fig. 3).

Characteristic mid-glacial formations, well-identifiable in several Hungarian loess profiles, are the forest steppe soil BD_2 and the steppe soil BD_1. Its cave counterpart is between the cave soil or soil sediment of sedimentation phases XIII-XV and the deposits from sedimentation phases X-XII correlated with loess l_4 appearing as a thin stadial formation (sedimentation phase XIV).

The middle glacial phase between the paleosols BD_1 and MF_2 is represented by the loess l_3 with intercalated geliflunctional-cryoturbational loess varieties, soil sediments and dells. It its cave counterpart (sedimentation phases XVI-XXIII) three cave soils or soil sediments can be identified. Under the cold and humid conditions of the intermediate stadials strong gelifluction and cryodeformation were active also in cave deposits (Fig. 3).


LITHOSTRATIGRAPHY
Relative Chronology
Loess
Cave Deposits
Lithology
1
2
3
4
5
Lithostrat-igraphical types
6
7
8
9
10
11
12
Upper Pleistocene
Late Glacial
Middle Glacial
Early G.
Eemian
Dunaújváros-Tápiósüly (5-10m)
Mende-Basaharc (20-25m)
2,0m-3,0m
13,0 m — 17,0 m
4,0 m - 6,0 m
5s. MF1
5fs. MF2
5s. BD1
5fs. BD2
5s. BA
3fs. MB1
2f. MB2
9H es.1
9H es.2
ss.
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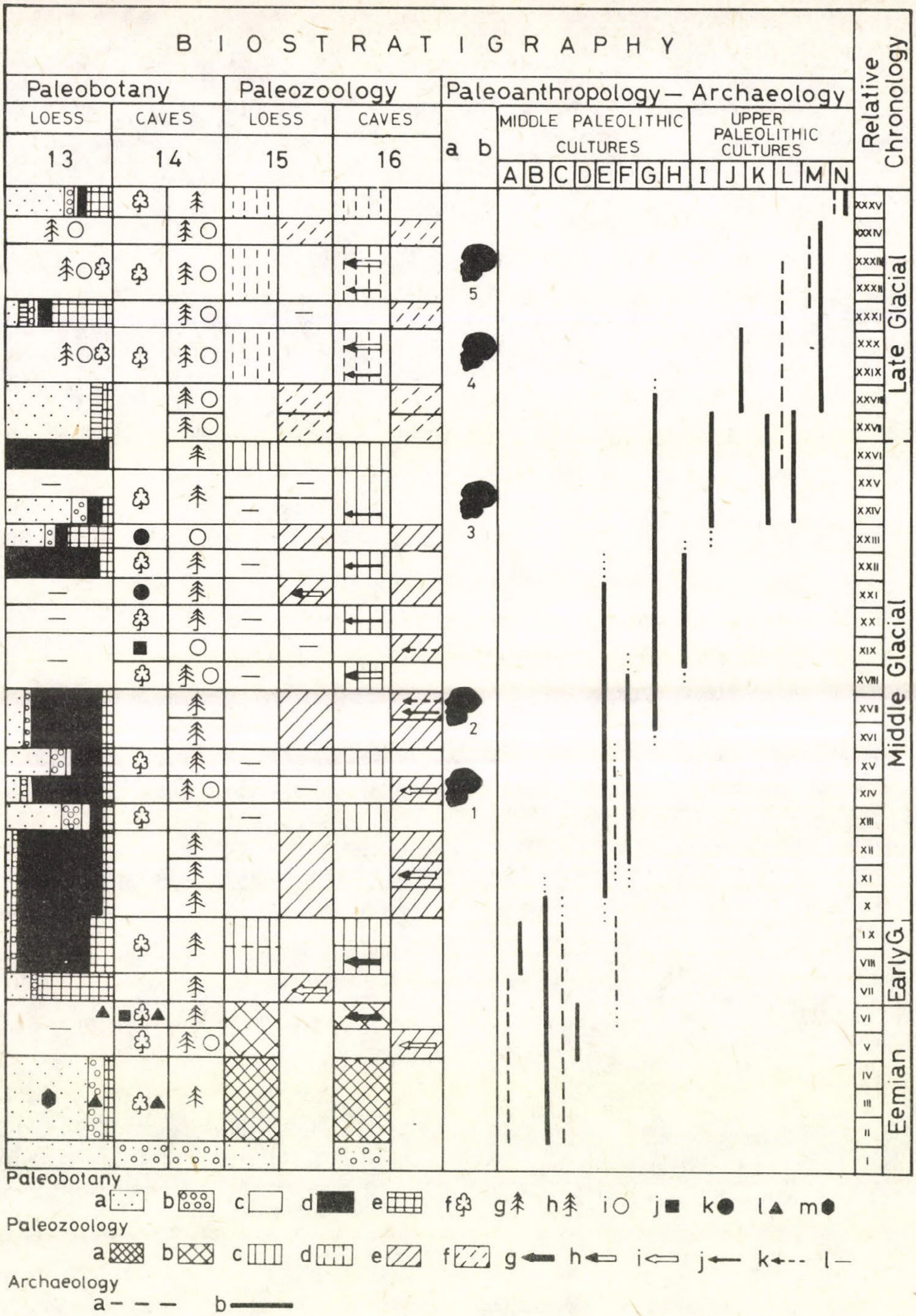
BIOSTRATIGRAPHY
Paleobotany
Paleozoology
Paleoanthropology — Archaeology
Relative Chronology
LOESS
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LOESS
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13
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MIDDLE PALEOLITHIC CULTURES
UPPER PALEOLITHIC CULTURES
A B C D E F G H I J K L M N
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Archaeology
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Fig. 3: Chronostratigraphical system of Upper Pleistocene loess and cave deposits in Hungary.

Lithostratigraphy

Loess

For lithology see Fig. 2. 1 = Series, 2 = Major periods of loess accumulation, 3 = Major periods of gelifluction, 4 = Major periods of erosion and dell formation, 5 = Cryoturbation, frost wedges, frost cracks.

Cave Deposits

For lithology see Fig. 2. 6 = Sedimentation phases, 7 = Total thickness, 8 = Major sedimentation phases: a = Frost shattered debris with many large blocks; b = frost shattered limestone with medium or small size particles, 9 = Major phases of chimney and debris fan formation: a = opening and expansion of chimneys with rapid debris fan accumulation below chimeys, b = slow expansion of chimneys and moderate debris fan building below chimneys, 10 = Cryodeformation, gelifluction: a = gelifluction + cryodeformation, b = gelifluction + strong cryodeformation, 'concassé' retouching on Paleolithic artefacts, c = moderate gelifluction, d = secondary (Dryas III) cryodeformation, 11 = Major periods of allochtonous loess accumulation, 12 = Erosion of layers.

Biostratigraphy

Paleobotany (13-14)

13 = Palynological data: a = indifferent wooden taxa: Pinus, Picea, Betula, Salix, Alnus; b = thermophilous wooden taxa: Quercus, Ulmus, Fraxinus, Acer, Corylus, Carpinus, Fagus, Juglans, Abies; c = Gramineae and Cerealia-Type; d = Artemisia, e = other herbaceous. 14 = Anthracotomic and fruit remnant investigation data. Dominant tree species: f = deciduous trees; g = Pinus; h = Larix-Picea, i = Pinus cembra. Less frequent tree species: j = Fagus; k = Quercus; l = Celtis; m = Vitis.

Paleozoology (15-16).

Last interglacial species occurring in both loess and caves: a = Sus scrofa, Capreolus, Dama sp., Ursus arctos, Asinus hydruntinus, Microtus arvalis, Myodes sp., Apodemus sylvaticus, Talpa europaea, Helicigona banatica; b = Sus scrofa, Capreolus, Ursus arctos (aff. tabubachensis), Microtus arvalis, Myodes sp., Apodemus sylvaticus, Talpa europaea, Spalax leucodon, Helicigona banatica.

Last glacial mammals occuring in both loess and caves:

Early and Middle Glacial	Late Glacial
Interstadial species:	
c = Sus scrofa, Capreolus, Alces, Ursus arctos, Talpa europaea	d = Sus scrofa, Capreolus, Ursus arctos, Talpa europaea
Stadial arctic-subarctic and steppe species:	
e = Rangifer tarandus Mammuthus primigenius (frequent), Equus big species, Microtus gregalis	f = Rangifer tarandus (frequent), Mammuthus primigenius (infrequent), Equus small species, Microtus gregalis

The most important immigrated small mammals:

g = Hystrix vinogradovi + Spalax leucodon; h = Dicrostonyx torquatus + Microtus gregalis; i = Lagurus lagurus; j = Lepus timidus immigration; k = Arvicola terrestris dominance pointing to high humidity, l = No or not studied data.

Paleoanthropology - archaeology

Anthropological finds: a = Homo neanderthalensis: 1 = Subalyuk cave, Subalyuk- type Charentian; 2 = Remete Upper cave, Jankovichian, b = Homo sapiens fossilis: 3 = Istállóskő cave, Aurignacian II; 4 = Pilisszántó niche I, Pilisszántóian, 5 = Balla cave, Gravettian

Archaeological cultures:

A = Tata- type Moustérian; B = Central European typical Moustérian, C = Central European Micoquian: Bábonyian; D = Central European Micoquian: Jankovichian; E = Érd- type Southeast- European Charentian; F = Subalyuk- type Charentian (RINGER, 1986); G = Early Szeletian in Bükk Mountains; H = Aurignacian I in Bükk Mountains; I = Upper Szeletian, J = Solutréo-Szeletian (RINGER, 1986), K = Aurignacion II in the Bükk Mountains, L = Eastern Gravettian, M = Pilisszántóian, N = Epipaleolithic, a = Open-air stations, b = Cave stations.

The paleosols MF_2 and MF_1 closing the middle glacial in the loess sequences of Hungary can be correlated with the interstadial double soil or soil sediment of sedimentation phases XXIV and XXVI. The generally arid late glacial conditions characterized by increased wind action are responsible for the important role of allochtonous sandy loess and loessy sand, also a main component of cave loess formed in karst hollows of elevated mouth, in the cave deposits correlating with the Dunaújváros-Tápiósüly loess series.

Significant accumulation of frost shattered debris only took place (after the phase XXVI) in the caves simultaneous to the evolution of humus horizons H_2 and H_1 intercalated into loess horizons l_2 and l_1, in the humid 'interstadials' around maximum glaciations (sedimentation phases XXIX-XXX and XXXII-XXXIII - Fig 3). The soils from Bölling and Alleröd interstadials in subaerial sediments (and of late glacial - BORSY et al. 1982) may correlate with the poor rendzina-like cave soil or soil sediment of sedimentation phase XXXV.

Biostratigraphy

In accordance with the major objective of this paper, the lithostratigraphical-chronological correlation of young loess and cave deposits, only some related paleobotanical-paleontological findings are treated. The most important data are tabulated (Fig. 3).

Paleobotanical data

Palynological-anthracotomic investigations (ZÓLYOMI, JÁRAI-KOMLÓDI, STIEBER, SKOFLEK, PASHKEVICH and URBAN 1953-1984) indicate that the temperate vegetation of the last interglacial also including mediterranean-submediterranean plants (*Celtis australis*, *Vitis*, *Juglans*, *Quercus pubescens*) was replaced by taiga-tundra vegetation during the last glacial as deciduous species gradually declined. In the upper glacial arctic-subarctic conifers (*Larix-Picea*, *Pinus cembra*, *Juniperus*) are predominant.

In the caves of medium-height mountains exclusively conifers are found for the periglacial periods (phases parallel with loess l_3) and mixed coniferous and deciduous forests in the interstadials (STIEBER, 1957).

The lower-lying loess areas were steppes with groves mostly of conifers (pines). In the interstadials the groves extended in area and were enriched in thermophilous and hydrophilous deciduous trees (JÁRAI-KOMLÓDI, 1968; URBAN, 1984).

Despite topographic and altitudinal differences, steppe and medium-height mountain vegetations are rather homogeneous during the formation of BD_1-BD_2, MF_1-MF_2 and the loess packet l_3 and the corresponding cave deposits.

The cold and humid climate of the accumulation of the loess packet l_3 is indicated by the appearance of *Fagus* and *Quercus* in the *Larix-Picea-Pinus* forests and by the high proportion of tree vegetation on lower-lying loess steppes (STIEBER, 1957; URBAN, 1984).

In both cases, for the soils BD and MF, the expansion of deciduous forests is documented below the lower members, BD_2 and MF_2 (STIEBER, 1957; URBAN, 1984).

Paleozoological data

The paleontological research on Upper Pleistocene climtic changes (MOTTL, KRETZOI, JÁNOSSY, KORDOS, VÖRÖS, KROLOPP, WÁGNER, 1938-1987) indicate a similar picture to vegetation reconstruction, from temperate with submediterranean-mediterrenean influence to arctic-subarctic climate.

Along with lithological and botanical data, the alternation of interstadials and periglacials is well indicated by proliferating temperate or arctic-subarctic mammal species. Among small mammals, the expansion of the arctic *Discroctonyx torquatus* and *Microtus gregalis* especially well correlate with cold maxima. The great dominance of the accompanying *Arvicola terrestris* is characteristic in the cold and humid phases of loess l_3 and H_1-H_2 formation (Fig. 3).

Archaeological and paleoanthropological data

The archaeological finds in the loess sequences and cave deposits of Hungary referred into the Upper Pleistocene blong to the following cultures: the two facies of the Middle Paleolithic-Central European Micoquian: Ábonyian and Jankovichian, Central European Moustérian, Tata-Type Moustérian, Érd-type Southeast European Charentian, Subalyuk-type Charentian and Early Szeletian; Upper Paleolithic Late Szeletian and Solutreo-Szeletian, Aurignacian I-II, Eastern Gravettian and the Pilisszántó cultures (GÁBORI, GÁBORI-CSÁNK, VÉRTES, T. DOBOSI, RINGER, 1953-1986). The archaeological cultures are associated with Homo neanderthalensis and Homo sapiens fossilis as attested by remains from caves (THOMA, 1968; GÁBORI-CSÁNK, 1983). The chronology of these cultures is shown in Fig. 3.

Among the archaeological cultures the Central European Micoquian Bábonyian facies and the Tata-type Moustérian contemporaneous subaerial and cave occurrences are most important for correlations between young loess and cave deposits during the last interglacial and early glacial (RINGER, 1983, 1986).

During the formation of the Dunaújváros-Tápiósüly loess series and the parallel deposition in caves in sedimentation phases XXVII-XXXV in the late glacial, simultaneous settlements of the Pilisszántó and Eastern Gravettian in the loess and cave Paleolithic localities (GÁBORI, 1964-1984, RINGER, 1986, T. DOBOSI et al. 1983).

SOME IMPLICATIONS FOR EUROPEAN CORRELATION OF COMPLEX LOESS AND CAVE CHRONOSTRATIGRAPHY

Some opportunities for the European correlation of Upper Pleistocene loess and cave stratigraphies (Figs. 2-3) based on litho- and biostratigraphic data and absolute dating are shown in Fig. 4. The long-range correlation of last interglacial and interstadial formations for the last 30-45 ka relies on C^{14} and TL datings, while for the period to the last glacial formations the TL data from the Mende and Paks key sections were used (BORSY et al., 1979; BUTRYM-MARUSZCZAK, 1984). We also applied the recent oxygen isotopic curve drawn from boreholes in the western Mediterranean (LABEYRIE, 1984 based on PATERNE). It allows correlations with the W-European Upper Pleistocene chronostratigraphy placed into EMILIANI's oxygen isotopic chronological system and with the best elaborated French cave chronostratigraphy (LUBLEY et al. 1972; LAVILLE, 1975). According to this attempt at long-range correlation the erosion section at the base of the Upper Pleistocene series in the Hungarian loess and cave sedimentation phases I-IV can be identified with the stage 5e by EMILIANI; the first interstadial of early glacial, the upper, forest steppe soil of the MB paleosol and cave sedimentation phase VI with the stage; the second early glacial interstadial, the paleosol BA and the cave sedimentation phases VIII-IX (cave interstadial formations) with the stage 5a (Fig. 4).

The loess l_4 of middle glacial and the cave stadial sequence of the sedimentation phases X-XII correlate with stage 4 by EMILIANI, i.e. the first cold peak of last glacial dated to 5ca 70 ka BP.

The paleosols BD_2 and BD_1 and the correlated cave sedimentation phases XIII-XV are identified with the double interstadial oscillation of EMILIANI's stage 3 (ca 60 ka BP - KORDOS, 1986; RINGER, 1986).

The paleosols MF_2 and MF_1 closing the middle glacial and the Mende-Basaharc loess series and the corresponding cave soils or soil sediments (phases XXIV and XXVI) belong to stage 3 by EMILIANI (the oscillations around 30 ka BP).

The Dunaújváros-Tápiósüly late glacial young loess series and the corresponding cave sediments deposited during the phases XXVII-XXXV are contemporaneous with EMILIANI's stage 2 and the cold maximum of last glacial.

CONCLUSION

In the paper attempt has been made - in the framework of modern, comprehensive research of cave sediments in the Geographical Research Institute Hungarian Academy of Sciences (PÉCSI - RINGER, 1986) - at correlating Upper Pleistocene loess and cave deposits supplemented with possible correlations with young loesses in Europe (PÉCSI, 1978) and with oxygen isotopic chronology.

Some paleogeographical and paleo(human)ecological conclusions for the Upper Pleistocene in Hungary drawn from abundant litho- and biostratigraphical data:

During the last interglacial (EMILIANI's stage 5e) a submediterranean-mediterranean influence is felt in soil formation, vegetation and fauna. At the same time, the introduction of Middle Paleolithic cultures (Tata-type Moustérian, Central European typical Moustérian, and Érd-type Southeast-European Charentian) seems probable arriving from the Mediterranean.

In the early glacial (stages 5d-5a by EMILIANI) ccol temperate climate became dominant as attested by soil formation, vegetation and faunal composition. This continued to the stage 4 by EMILIANI, a cold peak ca 70 ka BP, when the local development of most of the archaeological cultures seems to end.

In the interstadials of the middle glacial, forest steppe and soils formed under the mixed grove forests of deciduous species and temperate steppes of low hill regions and lowlands. During the stadials arctic-subarctic elements prevailed in both flora and fauna.

The stage 3 by EMILIANI is the main period of Ursus spelaeus hunting, numerous occasional hunting settlements are found. The relations of archaeological cultures are mostly with areas on similar latitudes (Jankovichian, Subalyuk-type Charentian and Early Szeletian). During EMILIANI's stage 2,

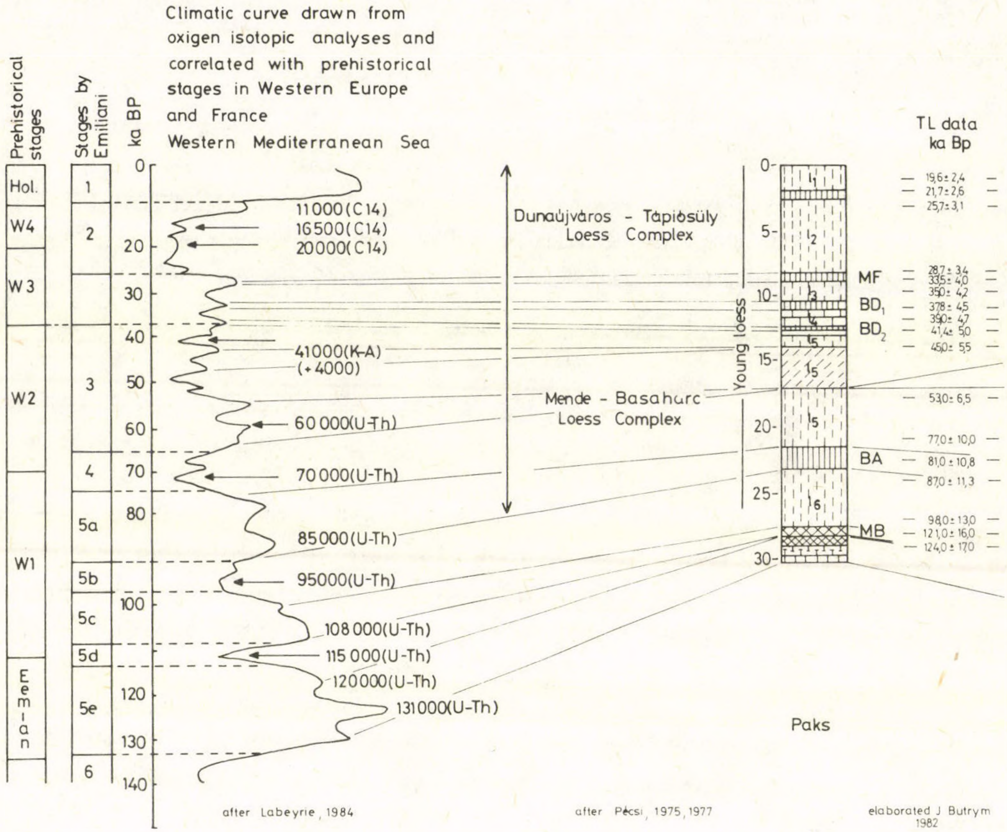
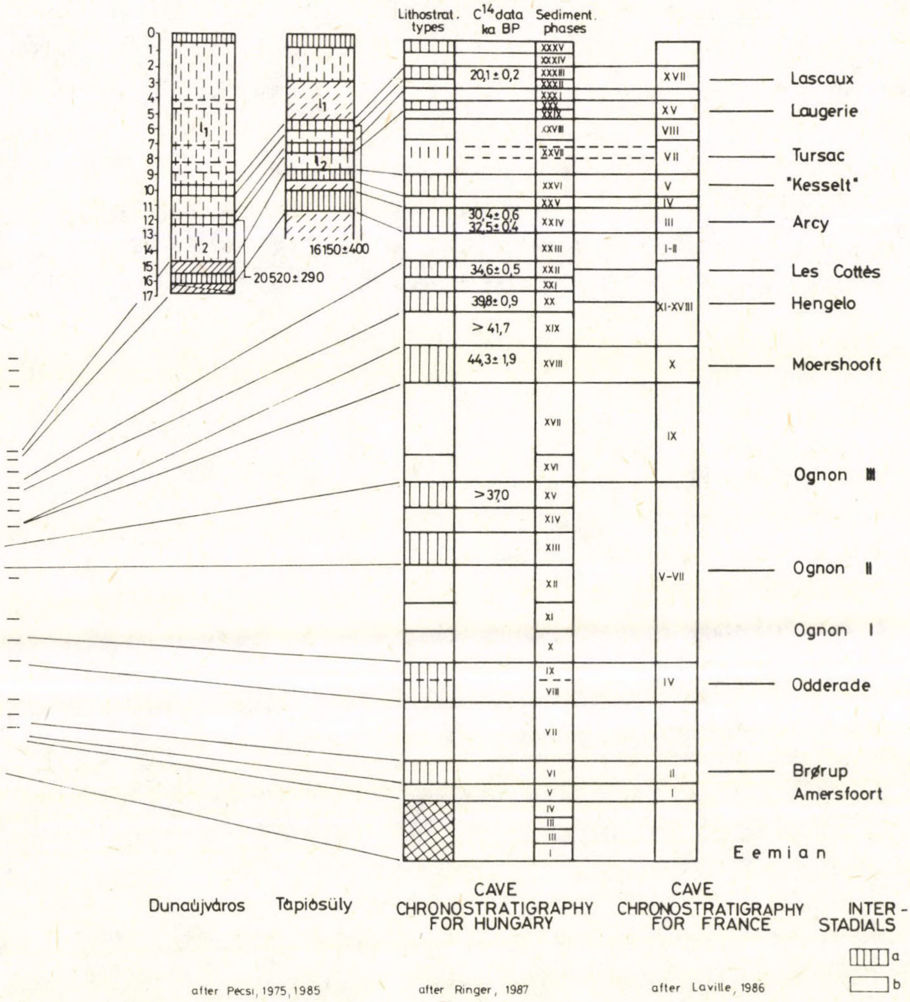


Fig. 4: Some aspects of European correlation of Upper Pleistocene loess - cave deposit chronostratigraphy for Hungary
a = interglacial and interstadial cave deposits; b = stadial cave deposits


Lithostrat. types
C14 data ka BP
Sediment. phases
20,1± 0,2
30,4± 0,6
32,5± 0,4
34,6± 0,5
39,8± 0,9
> 41,7
44,3± 1,9
> 37,0
16150±400
20520± 290
Lascaux
Laugerie
Tursac
"Kesselt"
Arcy
Les Cottès
Hengelo
Moershooft
Ognon III
Ognon II
Ognon I
Odderade
Brørup
Amersfoort
Eemian
Dunaújváros
Tápiósüly
CAVE CHRONOSTRATIGRAPHY FOR HUNGARY
CAVE CHRONOSTRATIGRAPHY FOR FRANCE
INTER-STADIALS
a
b
after Pécsi, 1975, 1985
after Ringer, 1987
after Laville, 1986

under cold and generally late glacial climate, the Hungarian Great Plain and the lower pediments of the Hungarian Mountains were covered by loess steppe with minor gallery forests.

With the disappearance of Ursus spelaeus an ecological crisis followed. The new trend in hunting was controlled by the proliferation of Rangifer tarandus and the small Equus caballus contained in the 'rein horizon' (PÉCSI, 1975).

The relations of the Upper Paleolithic Eastern Gravettian and Pilisszántó cultures are in northern, norhtwestern and northeastern directions, probably reflecting special transhumance adjusted to the seasonal wandering of reindeer herds (GÁBORI, 1964-1984; RINGER, 1986; T. DOBOSI et al. 1983).

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ECOLOGICAL CHANGES IN THE TERRITORY OF HUNGARY DURING THE HOLOCENE

S. Somogyi

ABSTRACT

Author has summarized the results of his investigations into the Holocene, the advent of human intervention into nature, in the treatise entitled 'A magyar nép kialakulásának és honfoglalásának környezete' /The geographical environment of development of the Hungarian people and the conquest of the Carpathian basin/ /SOMOGYI, 1983/.He analyzed the changes in the ecological factors /tectonic movements, climate, fluvial action etc./, their significance in controlling the environmental conditions of society. Recent geomorphological, litho-, bio-, and chronostratigraphical results based on modern absolute dating procedures /such as dendrochronology/ and relative dating /such as palynological analyses and archaeological topographies/ serve to make reconstructions of drainage and vegetation changes and the population history of lowlands.

INTRODUCTION

Geography can only fulfil its task if describes the phenomena and processes of the environment and if investigations are made in cooperation with other natural and social disciplines. In the evolution of the human environment special attention has to be paid to recent periods, particularly to the Holocene, when the physical environment followed a natural course of evolution which was increasingly affected by human intervention.

The treatise entitled 'A magyar nép kialakulásának és honfoglalásának földrajzi környezete' /The geographical environment of development of the Hungarian people and the conquest of the Carpathian basin - SOMOGYI, 1983/ summarizes the investigations into the paleoecological changes during this period. The present paper focuses on statements relevant to the territory of Hungary.

The leading principle of study was to estimate the impact of changes in ecological conditions on the potentials of human life. The recent achievements in geomorphology, litho-, bio-, and chronostratigraphy - significantly promoted by modern absolute dating procedures /such as dendrochronology/ and relative dating /such as palynological analyses/ - are increasingly reliable in both space and time. Archaeological research has provided an especially good starting-point, since the pattern of well-defined and well-described sites visually demonstrated the suitability of the individual regions for human settlement under the conditions of the time.

Tectonic movements

Altough the extent of Holocene crustal movements is much behind the figures for the Pleistocene /the estimations of M. PÉCSI - confirmed by several data - indicate $\pm$ 10-20 m for the Danube valley/, their consequences in hydrography and other physical elements are remarkable. Negative movements are located along the margins of lowlands and basins. They have influenced the courses of rivers /the Danube, Tisza, Szamos, Körös and Zagyva/ considerably along certain sections. The surfaces of young subsidence lack Paleolithic and Mesolithic finds. Uplifts also took place, the most important of them affecting the Nyírség region. This movement forced the Tisza river to follow the present course between the present settlements of Vásárosnamény and Tokaj. Tectonic movements made the Danube build its vast alluvial fan in the Little Plain /SCHMIDT-ZÓLYOMI, 1940; PÉCSI, 1959; SOMOGYI, 1973 - see Fig. 2 p. 116 in LÓCZY, this volume/.

CLIMATIC OSCILLATIONS

The changes of climate during the Holocene are generally placed on time-scales set up by the various, mostly vegetation chronological, methods elaborated in Northern and Western Europe. The geographical location of Hungary, its position relative to factors of climatic control in the narrower or broader environment /oceans and mountains/ create conditions which are different from those in the abovementioned parts of the continent. The main difference is that while Holocene climatic oscillations in the north are manifest in temperature curves parallel with the radiation curve, in Hungary a more complicated situation exists and for precipitation conditions the picture is even more varied. The estimated duration of climatic phases and their boundaries also show variation. In general, it can be claimed that Atlantic influence reached Central Europe later and stopped earlier than in Northwest-Europe /SOMOGYI, 1962; JÁRAI-KOMLÓDI, 1966; KORDOS, 1982/.

In Hungary settlement in the lowlands dates back to the Boreal phase /from 10,000 B.P./. Human settlements from the Atlantic phase /7000-5000 B.P./ are more frequent. The expansion of forests put an end to moderate alluviation of rivers characteristic in the Boreal phase. Instead, river mechanisms tended to produce large meanders. The resulting natural levees, point bars and riverbank dunes provided surfaces for the later developing classical Neolithic sites. They are primarily fishermen's and hunters' sites. On loess-mantled surfaces, however, stock-breeding prevailed and this occupation prevented the reforestation of the landscape /KRETZOI, 1977/. No finds have been recovered from sand surfaces, since Boreal blown-sand movements provided unfavourable living conditions /SOMOGYI, 1971, T. DOBOSI, 1975; KROLOPP, 1977/.

RIVER TERRACES

Along the rivers that kept their courses Preboreal incision into the valley floor accumulated during the Upper Pleistocene resulted in the formation of the first flood-free terrace, lowest and closest to the river channel. Due to its location related to living channel, this terrace was of outstanding importance in the allocation of settlements /PÉCSI, 1959; SOMOGYI, 1962; BORSY-FÉLEGYHÁZI, 1982/.

Holocene terraces or higher flood-plain levels were formed in the Subboreal phase /5000-2800 B.P./. This youngest incision phase of rivers indicates abundant water. The contemporaneous intensive fluvial erosion is evidenced by numerous finds of logs, dendrochronologically dated, recovered from Danubian deposits in the Alpine foreland. Dwellers of the Neolithic sites had to leave their homes as a result of flooding prior to the advent of river incision. As a consequence, there was a reduction in the number of sites in Transdanubia and the Great Hungarian Plain during the Copper Age compared to the previous age. It is of general cultural historical importance that cereals appear in mass in the pollen spectra from this phase. Also in Copper and Bronze Age sites the remains of domesticated animals became prevalent among animal bones. Although, as a result of increasing cultivation and stock-breeding, forests did not expand to the maximum extent motivated by physical potentials, in some marginal parts afforestation was considerable. In such areas Boreal chernozems turned into brown forest soils during the Subboreal /SOMOGYI, 1962; BECKER-FRENZEL, 1977/.

SETTLEMENT IN COPPER AND BRONZE AGES

During the Copper and Bronze Ages in mountains and hills people settled along rivers and in small basins. In lowland regions settlement is densest on flood-free surfaces, particularly along the margins of loess-mantled, higher-lying plains. Conclusions can be made from the physical conditions to differences in ways of life. The altitudinal distribution of sites is demonstrated by the following figures. In the Neolithic settlement reached up to 300 m above sea level in the medium-height mountains, while in the Bronze Age they reached 400 m on the average and locally occurred even at 800 m above sea level. New aspects were added to the selection of site. While in the Neolithic Age accessibility was prevailing consideration, in the Copper Age rising flood levels forced people to withdraw to higherlying surfaces. With the tribal warfare of the Bronze Age, defendability also became an important factor. Fot this reason flood-free islands and steep elevations /for instance, the Jakabhegy at Pécs/ became populated.

DRAINAGE SYSTEM IN HISTORICAL TIMES

The last period of natural drainage evolution took place in historical times /SOMOGYI, 1961/. The final formation of streams is reconstructed using a series of geological, sedimentological and mineralogical, geomorphological and potamological evidence supplemented by soil and phytogeographical data and archaeological finds. But written /or rather drawn/ evidence also exists in the form of Ptolemy's atlas from the 2nd century, where the areas of uncertain drainage in the Carpathian basin, different from the present, show the same large-scale hydrographical changes reconstructed by geological and geomorphological investigations /FRÖHLICH, 1885; SOMOGYI, 1971/. Even extended human interference can be traced /in the form of the Csörsz ditch/.

HUMAN INFLUENCE ON VEGETATION AND STREAMS

The present cultural pusztas of the Great Plain formed in the Subatlantic phase /from 2800 B.P./. They reflect human influence since the vegetation appropriate to climate would have been forested steppe. However, the first period of large-scale deforestation in Transdanubia also took place in the Iron Age /beginning at the same date/ and was due to the numerous Celtic-Roman settlements. The brown forest soil of the cleared forests in loess-mantled surfaces /e.g. the Mezőföld/ took an evolution trend towards the chernozem type. Deforestation and grazing locally induced smaller sand movements again and cover sand surfaces also came about. In river mechanism meandering /'middle reaches' type/ became general /with the exception of uplifting margins and the mountains/. This prevented the continuation of Subboreal incision and during low water stages they were meandering on the flood-plain confined within Holocene /No 1/ terraces and during flood stages running out to the surface of this terrace. The considerable decay of forests was caused by meandering as attested by the second erosion stage defined by dendrochronology /SIMON, 1957; SOMOGYI, 1962; JÁRAI-KOMLÓDI, 1966; ZÓLYOMI, 1980/.

The autodymanics of rivers produced cycles for each rivers, according to external influences. Along the Great Plain course of the Danube, for instance, the total cycle of meander evolution takes cca 200 years. For the Tisza river of slower flow this period is much longer, while, as a matter of course, it is shorter for rivers of higher gradient and more rapid flow /SOMOGYI, 1982/.

Although cultivation and grazing had transformed flood-free surfaces into cultural pusztas by the time of the Hungarian Conquest /895/ of the Carpathian basin, the landscape of cultural steppe was interrupted by marshes and swamps alternating with deciduous groves on the extended flood-plains /half of the present territory and one-eighth of historical Hungary/. The predominance of forested steppe, conforming to the given climatic conditions, was prevented by society's way of life. In suitable places, however, this vegetation did develop as it can even be seen from present-day reconstructions of paleovegetation. Where forest clearing affected previously developed oak forests, brown forest soils were replaced by chernozem brown forest soils /BALASSA, 1973; BÓNA, 1984/.

CHANGES IN LANDSCAPE TYPES

During the historical analyses of landscape types 10 types were identified within the lowland, hill and mountains main categories and they comprised altogether 21 typological units. In the order of use by society, first are flood-free loess-mantled plains, the marginal zones of which accomodated most of the winter shelters. In a more restricted number dwelling sites occur in forest clearances too. Even less frequent are the sites on lower hill levels, in valleys and basins of low mountains. On the higher flood-plain levels between forested steppes occupations as grazing only after annual floods and, in some places still affected by floods, fishing also occur. The otherwise abandoned lower flood-plain, levels, the marshes and swamps were winter shelters as well as the scenes for fishing, a wide-spread activity of the time /FODOR, 1942; KALICZ, 1957; PÁRDUCZ, 1941, 1959/.

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MAPPING RECENT GEOMORPHIC PROCESSES IN HUNGARY

Gy. Lovász

ABSTRACT

Author has elaborated a method to map the most frequent geomorphic process in a given area in its spatial extension. In the memoir subordinate processes are also indicated. The use of geological and hydrometeorological information is illustrated on an example.

INTRODUCTION

To study the geomorphic processes of historical times is a developing trend in geomorphology today. Social activities exert more and more influence on relief evolution and its dynamic equilibrium is often upset. As a consequence, undesiderable processes arise or the intesity of the previous ones grows multifold. The information collected during mapping is integrated into the investigations aiming at environmental protection and environmental imapct statement. When extending the concept of 'recent' times into the last 10,000-15,000 years, these investigations may also promote the knowledge on the geomorphological processes of the Holocene.

In Hungary mapping was initiated by PÉCSI, M., who presented the processes at small scale on the geomorphological map of the 'Atlas der Donauländer' (PÉCSI 1978). The elaboration of the methods for detailed mapping at large scale began in 1983.

OBJECTIVES

- Identification of areas with denudation and accumulation.
- Determination of the areas affected by different processes of erosion.
- Determination of the areas affected by different processes of accumulation.
- Estimation of the quantitative parameters of denudation and accumulation processes.
- Study of the singular fluctuation of present-day geomorphic processes.

METHOD

The separation of the areas affected by the particular types of processes is aided by frequent field observations (LOVÁSZ 1985). The observations in the periods of rainfall are especially important.

The interpretation of aerial photographs can also be of major help, since it primarily facilitates regional extrapolation.

In the judgement of the temporal dynamism of the types of processes the long-term hydrometeorological data are invaluable. In Hungary the analyses were carried out relying on temperature and precipitation data between 1901 and 1980. In order to reveal the singularities over several hundred years, glaciological, paleoclimatological, archaeological and other observations should also be applied.

RESULTS

The test area studied from 1983 has a surface of 850 km^2 with variable topography and lithology. Thus, 10 processes could be identified in space:

- intensive material transport by rainwater,
- intensive material transport by rainwater, with landslide hazard,
- moderate material transport by rainwater,
- very moderate material transport,
- deluvial accumulation,
- fluvial accumulation,
- lacustrine-paludal accumulation,

- deflation,
- neutral geomorphic evolution,
- processes characteristic of the built-up areas of settlements.

In a given area a wide range of geomorphic processes are active. The most active or the most frequent processes are mapped in their spatial extension. In the map memoir the secondary processes with subordinate role and lesser influence on geomorphic evolution are also indicated. The slope wash of various intensity (strong, restricted, highly restricted) is related to the slope length, inclination and crop pattern. Information on the changes of intensity is given in Tables 1 and 2.

Deluvial accumulation is typical on foot-slopes and valley margins. Fluvial accumulation is characteristic of flood-plains limited by flood condtrol dykes. Lacustrine-paludal accumulation is predominant in the undrained fluvial depressions. Neutral geomorphic evolution is indicated by the complete non-eroded soil profile.

In case of some arosive processes the singular changes of intensity are the functions of monthly (seasonal or annual) sums of precipitation. In N-Transdanubia the precipitation in July has grown between 1901 and 1980, while drier and wetter spells alternate (Fig. 1). This is an indirect indication of the increasing trend of sheet wash in N-Transdanubia in this century.

In its western part the Mecsek Mountains, S-Hungary, is built-up, by karstic rocks and neighbouring Pliocene clays. Part of the water percolates from the karst into the clayey and sandy layers. Winter precipitation (December to February) grew rapidly from about 1915 to 1960 (Fig. 2) and fell to our days. In theory it is justified to assume that in the mentioned periods karstification and, in the neighbouring clayey areas, landslide processes became intensified.

Table 1: Different intensities of areal slope wash in loess-covered surfaces, under the climatic conditions of Hungary

slope % / length m	arable	vineyard	meadow	forest
flat	0	0	0	0
0–5 % / 50 m	1	1	0	0
0–5 % / 50–100 m	2	2	1	1
0–5 % / 100 m	2	3	2	1
5–12 % / 50 m	2	2	1	1
5–12 % / 50–100 m	3	3	2	1
5–12 % / 100 m	3	3	2	2
12–17 % / 50 m	2	2	1	1
12–17 % / 50–100 m	3	3	2	1
12–17 % / 100 m	3	3	3	2
17–25 % / 50 m	3	3	2	1
17–25 % / 50–100 m	3	3	3	2
17–25 % / 100 m	3	3	3	2
25 % / 50 m	3	3	2	1
25 % / 50–100 m	3	3	3	2
25 % / 100 m	3	3	3	2

0 = neutral state; 1 = very moderate slope wash; 2 = moderate slope wash; 3 = intensive slope wash.

Table 2: Different intensite of areal slope wash on impermeable surfaces (of clay, mud, schlier etc.) under the climatic conditions of Hungary

slope % / length m	arable	vineyard	meadow	forest
flat	0	0	0	0
0–5 % / 50 m	0	1	0	0
0–5 % / 50–100 m	1	1	0	0
0–5 % / 100 m	2	2	1	0
5–10 % / 50 m	1	1	1	0
5–10 % / 50–100 m	2	2	1	1
5–12 % / 100 m	3	2	2	1
12–17 % / 50 m	1	1	1	1
12–17 % / 50–100 m	2	2	1	1
12–17 % / 100 m	3	3	2	1
17–25 % / 50 m	2	2	1	1
17–25 % / 50–100 m	2	3	2	1
17–25 % / 100 m	3	3	2	2
25 % / 50 m	2	2	2	1
25 % / 50–100 m	3	3	2	2
25 % / 100 m	3	3	3	2

0 = neutral state; 1 = very moderate slope wash; 2 = moderate slope wash; 3 = intensive slope wash.

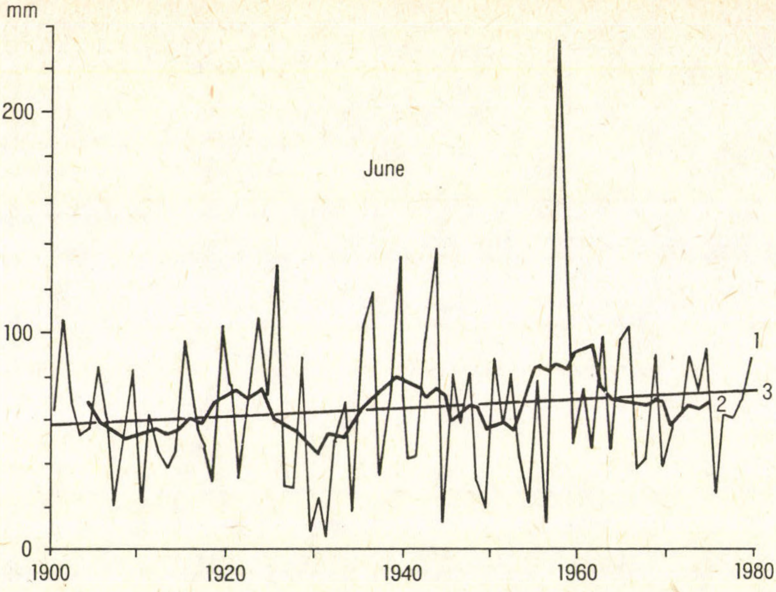

Fig. 1: Changes of total precipitation for June in Northern Transdanubia (station: Bábolna).

1 = monthly sums; 2 = moving trend; 3 = linear regression line

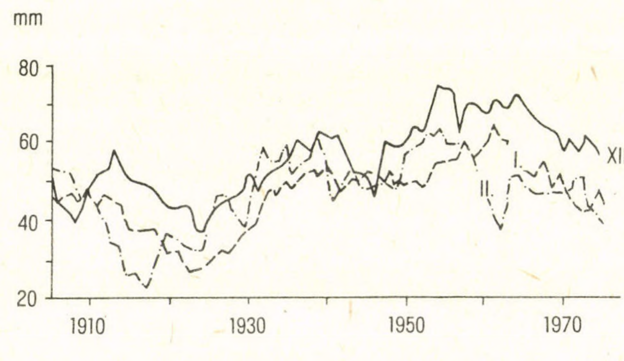

Fig. 2: Moving trend of the change of precipitation for December, January and February between 1901 and 1980 in the Mecsek Mountains (station: Abaliget)

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GEOMORPHOLOGICAL SKETCH OF THE NORTHERN LITTLE HUNGARIAN PLAIN

M. Kaiser

ABSTRACT

New geological, pedological, geophysical, paleontological and geomorphological evidence has accumulated for the two northern regions of the Little Hungarian Plain, the Szigeköz and the Győr–Tata terraced plain. Their evolution is described with special regard to the placers containing gold deposited by the Danube. Attempts are made at the absolute dating of various geomorphological levels.

INTRODUCTION

In the Hungarian State Geological Institute the comprehensive geological research of the Little Plain began in 1982. It involves the preparation of various geological, engineering geological, hydrogeological, pedological, geophysical and geomorphological maps at 1:100,000 scale. Most of the are based on field surveys and the interpretation of aerial photographs.

Geomorphological investigations began on the northern margin of the Little Plain. The western part is constituted of the elongated island between Danubian channels, the Szigetköz, and the eastern part is the Győr-Tata terraced plain.

The geomorphic action of the Danube in the area started in the Neogene. A vast alluvial fan accumulated over the subsiding surface from the west and it was building until the Middle Pleistocene (SZÁDECZKY-KARDOSS, 1938; PÉCSI, 1959, 1962; PÉCSI - KRETZOI, 1979). In the Middle Pleistocene the central and norhern parts of the fan were affected by renewed subsidence and the formation of a younger alluvial fan commenced. The surfaces not affected by this new subsidence were eroded by the Danube and terraces developed (PÉCSI, 1959). Today only small remnants of the old fan are detected on the Parndorf gravel plateau in Austria and as the highest terraces ot fhe Danube between Győr and Tata in Hungary. The younger fan extends from the national border to the town of Győr and no terraces were carved of its surface. Alluvial deposits of various age are found superimposed and juxtaposed to each other (KRETZOI - PÉCSI, 1982; PÉCSI et al. 1983; ERDÉLYI, 1979).

SZIGETKÖZ

It is part of the younger alluvial fan enclosed between the channels of the Danube (Fig. 1 - see gomorphological sketch in LÓCZY's paper). On the surface subsided in the Lower Pleistocene and in several phases since the Middle Pleistocene a thick alluvial sequence was deposited. Quaternary sediment thickness amounts to 100-400 m, underlain by Pannonian formations. The surface of the Szigetköz is exclusively of Holocene deposits.

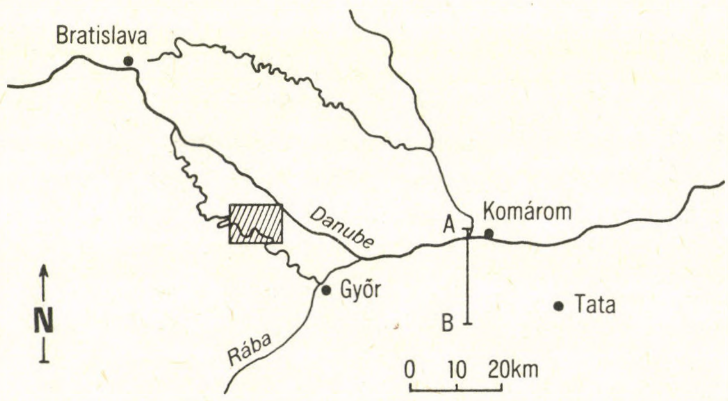

Fig. 1: The northern Little Plain with the area represented on the map of Fig. 2 and the A-B section shown in Fig. 3.

It is a flat lowland with very subdued (some metre) relief. Landforms are mostly of fluvial and subordinately of deflational origin. Before river regulation the main and by-channels of the Danube were shifting widely over the surface. The main channel was braided, while the by-channels were of meandering type. Shifting channels and wandering meanders produced multiple point-bar systems. Most of the surface exhibits a maze of meander remnants and point-bars, frequently overlapping or superimposed on each other (Fig. 2). The diameters of meander arcs are 500-1000 m along the main Danubian channel and the Moson-Danube, while this parameter has values of 200-500 m and sometimes 1000 m for smaller by-channels.

The Danube formed a lower and a higher flood-plain level in the Szigetköz. The relative height of the lower level is 1-2 m above the mean water of the river. It lies along the main channel and the lower Moson-Danube in a narrow belt. It can also be traced in smaller patches in the area of ox-bows to completely filled channels with excess water only appearing during floods (PÉCSI, 1959, 1968; GÖCSEI, 1979). The floors of ox-bows and channel remnants are 1-3 m below the general surface and have 2-3 m fills of silt and clayey silt. The higher flood-plain level is primarily built-up of shoal remnants, it has a relative elevation of 4-6 m. It is locally overlain by blown sand (N of Győr) and it means an additional 5-10 m elevation. Faunal investigations and radiocarbon dating indicate early Holocene age for this level (PÉCSI, 1959, 1974; PÉCSI et al. 1983).

Before river regulation gold-panning was a characteristic occupation in the Szigetköz. The concentration of heavy minerals is rather fluctuating both horizontally and vertically, on the average it is below the industrial level (PANTÓ, 1935). Heavy mineral occurrences are now being re-investigated for the exploration of other economic minerals. Studying one of the largest point-bars, PANTÓ confirmed the old observation that placers form along the upper diverging part of shoals. In 1986 two additional point-bars were sampled by drilling in regular pattern. The laboratory investigations are under way.

Point-bars and former channels are distinct in aerial photographs. The interpretations enabled me to show on the geomorphological map of the Szigetköz the areas where geomorphological position favoured heavy mineral placer accumulation (Fig. 2).

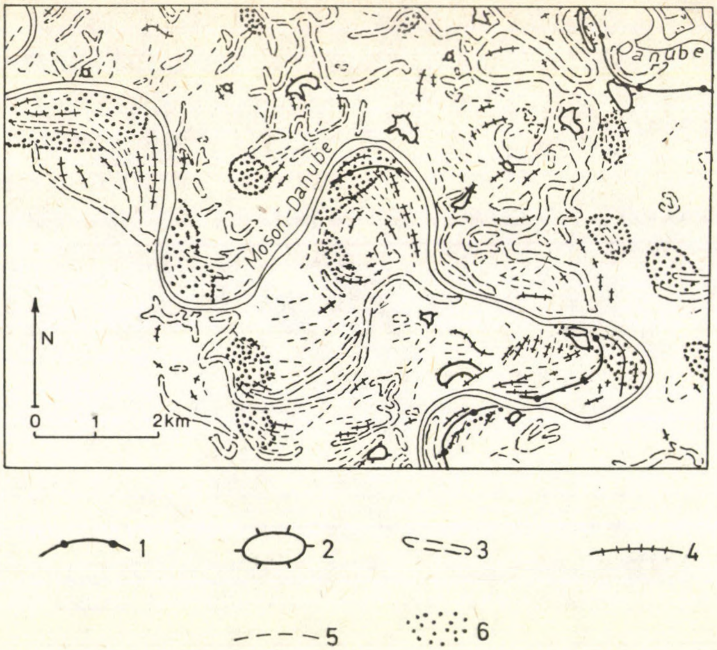

Fig. 2: Hypothetical distribution of heavy mineral bearing placers. Detail from the geomorphological map of the Szigetköz and environs. 1 = margin of higher flood-plain level; 2 = remnant of major shoal, 3 = ox-bow, filled channel; 4 = point-bar; 5 = depression between point-bars; 6 = sites with geomorphological position favouring the development of heavy mineral placers

GYŐR-TATA TERRACED PLAIN

The area belonged to the older Danubian fan and only was moderately affected by the subsidence continuing since the Middle Pleistocene. It has a surface of terraces arranged in a step-like fashion (Fig. 3).

The first terrace (no IIa in the Hungarian system) has relative elevations of 8-12 m. It is built-up of 4-10 m thick sandy gravel, mostly covered by blown sand. The age of the terrace is estimated from faunal finds and geomorphological investigations as the second half of the Würm (PÉCSI, 1959, 1974; FRANYÓ, 1971). The second terrace (no IIb) has relative heights between 18-22 m with 4-10 m of fluvial sandy gravel mantled by loess, sandy

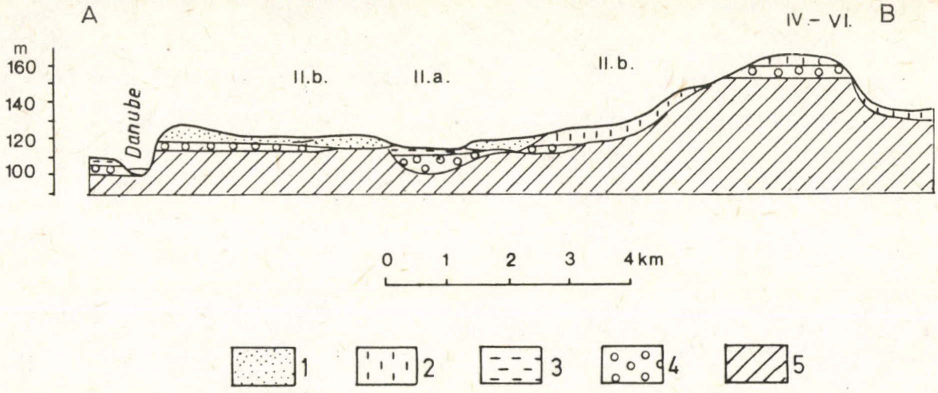

Fig. 3: N-S section of the Danubian terraces at Ács, W of Komárom. 1 = blown sand; 2 = loess and deluvium; 3 = fluvial silt; 4 = fluvial sandy gravel; 5 = Pannonian clay, clay-marl, aleurite and sand

loess and locally blown sand. Paleontological evidence, absolute dating and geomorphological data indicate that terrace formation lasted from the late Riss to the early Würm (PÉCSI, 1959, 1973). As determined in the Gerecse Mountains through the Th/U dating, the age travertine horizons covering the terrace are 70-100 ka B.P. (PÉCSI, 1973; PÉCSI et al. 1983). There are further scarps some metres above terrace no IIb. Attempts have been made to decide whether they are remnants of the poorly developed third terrace of the Danube or elevated parts of terrace no II/b (PÉCSI-KRETZOI, 1979; PÉCSI et al. 1983). The highest lying fluvial sandy gravel and gravel beds are remnants of the older Danubian fan (SZÁDECZKY-KARDOSS, 1938; PÉCSI, 1959). In PÉCSI's opinion the deposits include materials of terraces nos IV, V, and the layers are in normal stratigraphic position. This assumption was confirmed by E. KROLOPP's malacological investigations (BERNHARDT et al. 1974). The fluvial deposits are 3-6 m thick. Recently, it has been proposed that part of the gravel is much older than Quaternary. The lowerlying layers of different grain size are evaluated by PÉCSI (personal communication, 1984) as Pannonian deltaic gravels. The deposition of the overlying, coarser gravel may have been induced by the material transport of tributaries coming from the north. This is supported by the southerly dip of layers, observed in several places and the occurrence of large gravels (of 15-25 cm diameter) probably deriving from the Carpathians. The alluvial fan terrace is now located at the summit level of E-W residual ridge. Relative relief is 35-45 m in the west and 75-85 m E of the Concó creek. The relief is manifested in a scarp. From the analyses of gravel composition, it was assumed that they represent the same terrace (SZÁDECZKY-KARDOSS, 1938; PÉCSI, 1959). However, during the geological mapping of the area during the early 1970's, a SE-NW fault-line was identified from the shifting of Pannonian layers and surface relief and this fault cuts through the terrace where the scarps are seen (BERNHARDT et al. 1974).

Fig. 3 shows a section of the Danubian terraces at Ács, W of Komárom (section A-B in Fig. 1). Lateral erosion undermines the bank of Pannonian clayey silt and clay marl. Along the undermined bank, terrace no IIb deposits are exposed at 8-10 m elevation. Along this section terrace no IIa was formed following an older Danube channel some kilometres away from the present river bed. S of the remnants of alluvial fan terraces nos IV, V and VI, preserved on hilltops a basin of denudational origin is located. The alluvial fan terraces nos IV, V and VI developed during the Lower Pleistocene. The age of terrace no IV closing the sequence of the former alluvial fan is Mindel glacial, while the age of travertine on the terrace at Vértesszőlős is more than 350 ka B.P. (PÉCSI, 1973).

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HOLOCENE CHANGES OF A FLOODPLAIN LANDSCAPE IN THE LITTLE HUNGARIAN PLAIN

D. Lóczy

ABSTRACT

Geomorphological, palaeobotanical, archaeological and historical-geographical data have been analyzed to assess the utilization of natural resources in a typical Danubian landscape unit of 375 km^2 area. A major turning-point of the development of cultural landscape is seen in the largescale flood control measures of last century, when the higherlying parts were involed in intensive cultivation and the braided channels and enclosed wetlands were preserved more or less intact. Today the Szigetköz is doomed to change fundamentally through the environmental impacts of the Slovak – Hungarian Gabčikovo (Bős)-Nagymaros Barrage System on the Danube to be completed by 1990.

OBJECTIVES

The reconstruction of cultural landscape evolution is aimed at following the prolonged and intricate processes through which human society exploited and transformed the natural environment. One of the most difficult tasks in this respect is to find evidence about the advent of farming and forestry in a region. These forms of sustenance date back long before written history. Recent developments in the natural and social sciences, however, has brought about methods the joint application of which may help reveal prehistorical physical conditions and human activities. New radiometric dating techniques, pollen analysis in paleobotany, reconstruction of economies in archaeology, the investigation of the surviving elements of ancient occupations in ethnography /anthropology/, stratigraphical analyses in Quaternary geology and mapping of paleoecological conditions in historical geography belong here.

Research in this field can be useful for nature conservation and landscape planning, since the comprehensive picture integrated from the results of special investigations unambiguously present the ancient features of the landscape to be preserved in contrast with the elements alien to it. Suggestions for the utilization of the landscape can also be concluded from the historical evolution of ecological change.

Relying on information from the above sources and particularly from the map of landscape types and of present-day land use, this paper is meant to assess the extent to which human influence is responsible for the contemporary state of the environment in the Szigetköz and to pinpoint the periods when certain human activities induced major ecological change. Future prospects can also be evaluated with knowledge of the past.

GEOMORPHOLOGY

The area studied, the Szigetköz /'interfluve of islands'/ is a typical Danubian landscape in the Little Hungarian Plain, the Hungarian counterpart of the Csallóköz /Žitný ostrov/ in Slovakia /Fig. 1/. The area was affected by major natural changes over the last 10,000 years. The formation of the younger alluvial fan of the Danube /of sand and silt/ took place in the Holocene over the more extended gravelly Pleistocene fan of locally more than 400 m thickness /Fig. 2/. The subsidence of the Little Plain basin goes on with varying intensity. The altitudinal difference between the apex and the fringes of the alluvial fan, however, has not considerably exceeded 10 m over slightly more than 50 km distance and, thus, the lateral erosion of channels frequently changing their courses has produced a flat surface. In the particular cases it is not always easy to distinguish the lower flood-plain level from the higher one, since the relative relief is merely 3-4 m between them /PÉCSI, 1968; GÖCSEI, 1979;

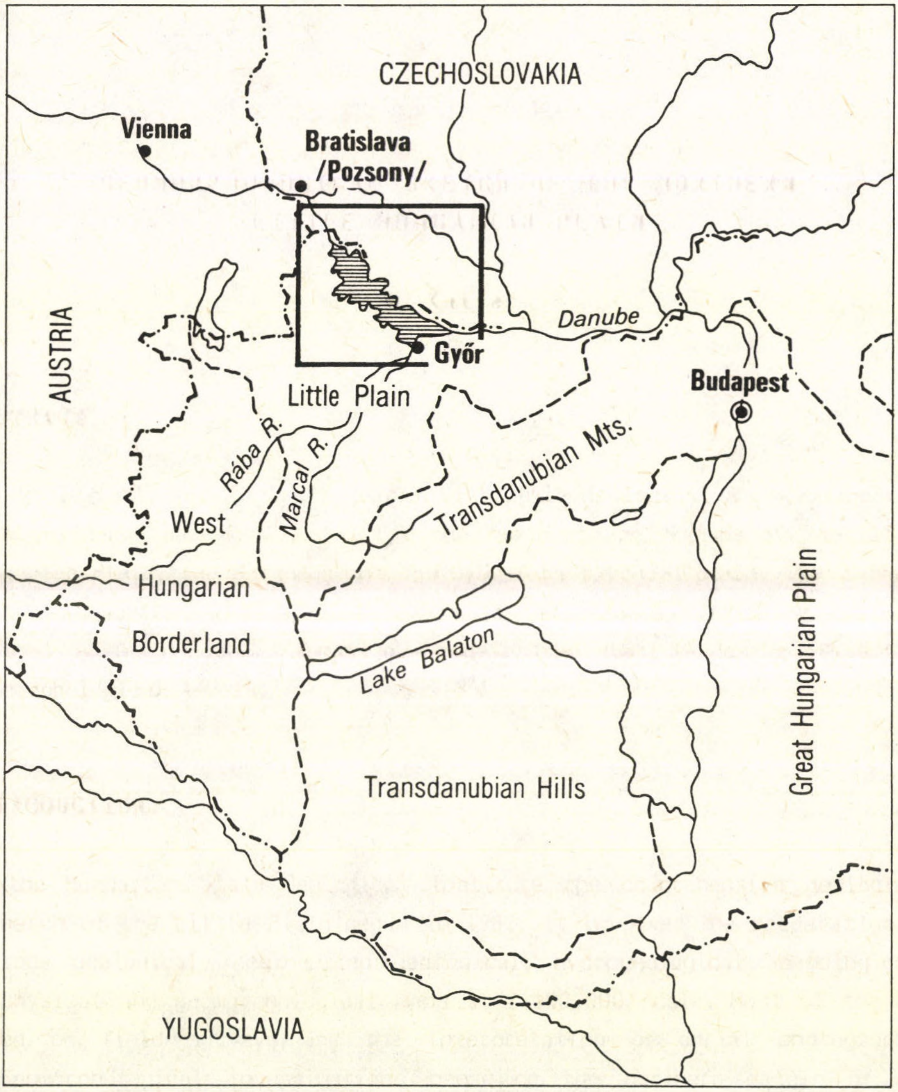

Fig. 1: Location of the Szigetköz in West-Hungary

ÁDÁM,-MAROSI, 1975/. Notwithstanding, in a lowland where groundwater conditions control habitats, small differences in elevation may be decisive ecologically.

On the map of landscape types in Hungary /PÉCSI - SOMOGYI, 1983/ the Szigetköz is shown as a partly flood-free /protected by dykes/ and partly seasonally inundated flood-plain on the alluvial fan with cut-off channels /bar-and-swale terrain/. It has alluvial, meadow alluvial, or swamp-forest soils mostly used for agriculture and a moderately warm and humid climate with subalpine and subatlantic influences. There are a few remnants of groves and marsh forests on the active flood-plain.

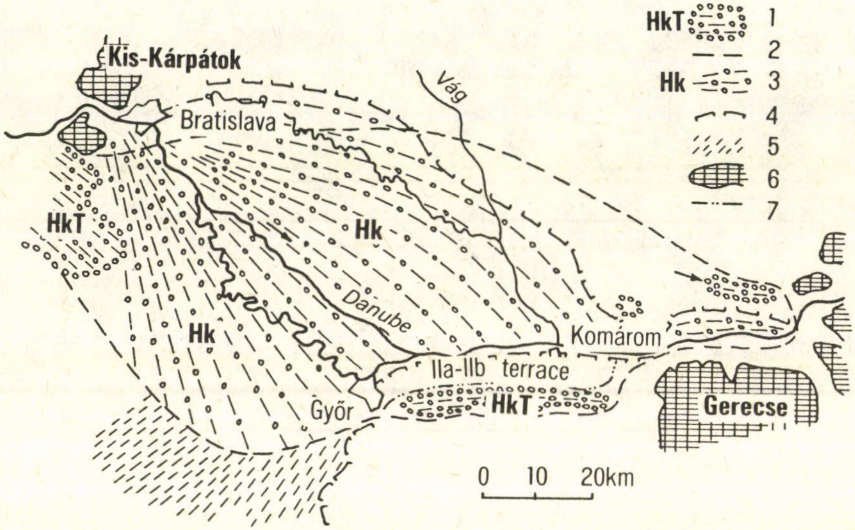

Fig. 2: The older and the younger alluvial fans of the Little Plain /PÉCSI, 1975/

1 = remnants of the older fan /HkT/; 2 = hypothetical extension of the early Pleistocene fan; 3 = extension of the Danubian alluvial fan accumulated to the Mindel glacial /Hk/; 4 = boundary of the younger alluvial fan accumulating since the Mindel-Riss interglacial; 5 = younger alluvial fans of the Rába, Répce, and Marcal rivers; 6 = mountains, 7 = river terraces

HISTORY OF HUMAN INFLUENCE ON THE LANDSCAPE

Prehistorical times

Since the subsidence of the area investigated accelerated in the early Holocene and it was partially inundated, geographers suggest that no finds older than Neolithic can be expected /SOMOGYI, 1984/. Among the braided channels which now enclose the Szigetköz, however, some sporadic finds have been recovered from the late Palaeolithic, including a stone-axe, indicate temporary settlement. Its density may have been the highest in the southernmost sandy parts of the area /Szitás heights, 123 m above sea level/. This part is also closest to the most important trading route of North-Transdanubia /the 'Amber Route' running north to south/ of later times /GÁBORI, 1964/. The first inhabitants were gatherers, hunters, and fishermen; their impact on the natural environment was presumably small and now untraceable. Although the warm and dry climate during Mesolithic times is regarded as favourable for human settlement, no finds have been found yet. The first agricultural activities in the environs of Lake Balaton, 120 km to the south of the Szigetköz were dated palynologically to 4500-4000 B.C. /ZÓLYOMI, 1980/, with small-size T r i t i c i u m pollen at the first date and large numbers of cereals from the Middle Neolithic. The easy communication between Little Balaton swamp, from where the pollen was washed into Lake Balaton, and the southern corner of the Szigetköz around Győr, along the Marcal river basin suggests that this dating is applicable to the latter region too. Apart from cereals, the expansion of F a g u s seems to be closely correlated with cultivation. The first expansion is in the younger Atlantic and the second, with C a r p i n u s , in the middle Subboreal, i.e. middle to late Bronze Age. As, at that time, cultural impacts were far from decisive in the landscape, the U l m u s decline is probably connected with climatic change rather than human activity. Early deforestations are indicated by minor rises in non-arboreal pollen. Archaeological finds for the Neolithic of northern Transdanubia /KALICZ, 1980/ suggest large villages with long houses. Slash-and-burn farming and forest cutting are assumed. Direct evidence for settlement in the Szigetköz is lacking. Floods of the shifting Danube channels on the alluvial fan might have destroyed settlements or their remnants /TIMAFFY, 1939/. The potential vegetation types /Fig. 3/ /elaborated from SIMON -

Fig. 3: Potential vegetation of the Szigetköz

1 = Fraxino pannonicae-Ulmetum; 2 = Salicetum albae-fragilis /triandrae; 3 = Convallario-Quercetum; 4 = Phragmition-Magnocaricion; 5 = built-up areas with ruderal vegetation; 6 = area inundated after barrage construction

HORÁNSZKY - KOVÁCS-LÁNG, 1980/ tell about natural associations and human interference can be evaluated through comparison with Fig. 4/.

Under the gradually deteriorating climate of the Subboreal, reduced carrying capacity only allowed animal husbandry as a main form of sustenance. In the Copper Age the Szigetköz was a sparsely populated area between two cultural groups. In the late Bronze Age, at cca 1100 B.C., the invasion of the ancestors of Illiric people took place from the west.

Historical times

The Celts were engaged in pig breeding, the Romans in viticulture, but not many details are known. The people of the Migrations /Longobards, Gepids and others/ were followed by the Avars in late 6th century A.D. The Avar economy has not yet been revealed in detail. Notwithstanding, the predominance of farming over stock-breeding /cattle, horses, pigs and poultry/ is probable in some periods.

Several impacts were manifest in the landscape when in 895 the tribes of the Magyar /Hungarian/ Conquest reached the area. These, however, were relatively insignificant compared to the changes brought about by the population expansion following the Conquest. The Magyars settled in the Carpathian basin during the so-called 'Little Climatic Optimum'. The Szigetköz, a pre-limes area during Roman times, was first used by the Magyars as a borderland /'gyepü'/ of swamp forests. The first written document, from 990, tells about the settlement of frontier guardsmen. The first village /Vének/ dates back to 1093 /TIMAFFY, 1980/.

Medieval history records several raids /Austrians in 1241, Czechs in 1271, Ottoman troops in 1529, 1594 and 1683/, but prolonged occupation was not achieved. The vicissitudes of history and, even more so, the repeated floods were effective in remodelling settlement pattern from time to time. For a long time, main occupation was fishing supplemented by gathering, hunting and gold-panning, activities with limited environmental impact. The early 19th century saw a peak of animal husbandry /partly fodder based/, when the forest-belt along the Danube and the parallel broad belt of meadows, narrowing down to the south, supplied large numbers of cattle, horses and pigs. Water birds were also important in the economy of this region with numerous ox-bow lakes. The hydrography before river regulation is shown in Fig. 5.

Fig. 4: Map of present-day land use in the Szigetköz /from topographic maps and field survey revised from LANDSAT TM colour composite for April 4, 1984/.
1 = arable land, 2 = forest, 3 = meadow and pasture, 4 = wetland /reed and sedge/, 5 = built-up area, gardens, orchards or vineyards. Barbed line shows flood-control dykes and pecked line marks the extension of the projected Hrušov /Dunakörtvélyes/ — Dunakiliti reservoir

Flood control was first decreed by King Sigismund in 1426, who defined it as a task of individual village communities /ALEXAY, 1982/. Modern river regulation /1880-1896/ was principally intended to prevent floods by accelerating runoff in channels between embankments /SOMOGYI, 1978/. In spite of the negative impact on river-bed sedimentation, the somewhat belated drainage measures and irrigation schemes gave rise to intensive farming in the Szigetköz too. The course of the Danube changed very little in the Szigetköz, because the flood-control dykes border a relatively wide active flood-plain with forests. Floods, however, are not yet events of the past. Even in 1954 a devastating one fed from meltwater and abundant summer rainfalls inundated the eastern half of the Szigetköz.

In the irrigated fields, Vienna horticulture proved to be profitable. Vegetable gardens /cabbage, potatoes/ and orchards /apples, plums and peaches/ are still wide-spread around villages. The utilization of thermal water from the Lipót borehole in hothouses promotes intensive gardening. The advent of commercial forestry was marked in the plantation of Populus euramericana cultivars for cellulose production.

Post-war social reforms had major impacts on the landscape. Collective farms, introduces in the 1950's and early 1960's, had an important ecological corollary: the small plots adjusted to the physical units, such as filled meanders or natural levees, merged into large fields comprising various habitats /ecotopes/. At the same time, the cooperative farms were able to use machines, fertilizers, and chemicals on broad scales to achieve higher yields.

The present use of the region /Fig. 4/ primarily reflects changes in hydrography due to human intervention during the last hundred years. Flood -control dykes appear as marked boundaries between forests /22 per cent of total area/ and wetlands /3 per cent/ of active flood-plains and agricultural land /arable 58 per cent/. Large-scale farming more or less obliterated the network of mostly filled meanders. Scattered remnants of the once contiguous pasture belt /7.5 per cent/ occur along the dyke.

Developments of the last decade, however, suggest that this zone should have had bright prospects /GÓCZÁN, et al. 1983/ if the Slovak-Hungarian Barrage System were not to be built.

The future

The projected scheme, intended to improve navigation on the Danube, is

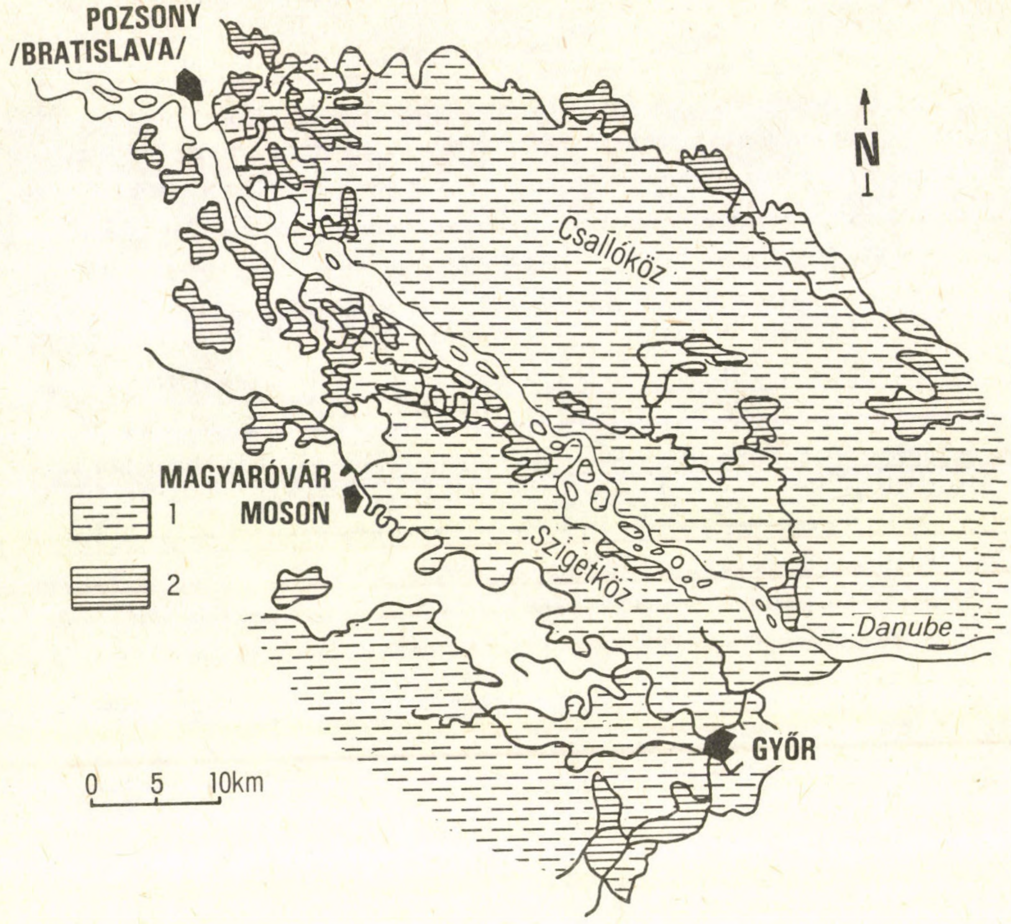

Fig. 5: Hydrography of the Danubian alluvial fan before last-century flood-control measures /from 'Magyarország vizborította...' 1938/
1 = areas inundated during floods, 2 = areas inundated over most of the year

expected to cause major changes in the environment. Most of the discharge of the Danube will flow through a diversion canal on the Slovakian side and the present active floodplain will have insufficient water to maintain its vegetation. The old channel will suffer major ecological damage affecting bank and aquatic biota. The pollution hazard of water reserves in aquifers will increase. The groundwater table will rise in the north around the projected reservoir and drop over most of the southern areas, thus influencing groundwater flow and producing an impermeable hardpan in the root zone of plants. Poplar stands may be reallocated and the crop pattern adjusted to the altered conditions. The meadows and pastures, which are in ecological balance today, will need irrigation or groundwater supply through the expensive infiltrating canal system.

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BIOSTRATIGRAPHIC INVESTIGATIONS IN A HOLOCENE BASIN OF TRANSDANUBIA

L. Füköh

ABSTRACT

Biostratigraphic investigations are significant contributions to Quaternary research in Hungary. The joint study of sediments in young basins and the fossil /mostly mollusc/ faunas contained in them allow paleoecological conclusions. Author attempts to reconstruct Holocene geomorphic and biosuccessional evolution in the Sárrét, Fejér county, the best studied of Holocene depressions in Hungary. He carried out sedimentological and faunal investigations on three sampling sites and from guide species established three stages in the evolution from a deep lake with open surface to a peat-bog. A relative correspondence of sedimentation history with climatic phases is attempted at.

INTRODUCTION

Late Pleistocene and Early Holocene subsidence allowed the formation of numerous extensive bogs /fens/ in lowlands of Hungary.

The Sárrét in Fejér county is the best-known region among them. First information on this region was given by T. KORMOS /1909/. He collected the malacological material first of all from the surface. Later E. KROLOPP /in: RÓNAI-SZENTES, 1972/ collected samples, focussing on the fauna of the subsurface sediments. Modern microstratigraphic-biostratigraphic investigations started in 1973 /FÜKÖH, L.-SZABÓ, I. 1975; FÜKÖH, L. 1976/. Meanwhile, peat and paludal lime excavation in the area allowed the pedological investigation of economic deposits /DÖMSÖDI, M. 1977/.

In this paper the mollusc fauna of the sediments and faunal changes are discussed. The two papers published earlier on this topic /FÜKÖH-KROLOPP, 1986/, the former with only partial results and the latter with a short summary. This paper is not meant to close the work at all; further investigations and the publications of manuscripts by several authors are planned.

DESCRIPTION OF AREA INVESTIGATED

The Sárrét of Fejér county is the largest depression in the Mezőföld loess plain . It has an area of 120 km^2, 20 km length and its width varies between 4 and 8 km. The formation of the basin started with Pleistocene /Riss-Würm/ subsidences of NW-SE and NE-SW axis /BULLA, 1959/. This is supported by the borehole profiles published by T. KORMOS /1909/ from the Pét and Gusztuspuszta region where the Pannonian clays are overlain by Upper Pleistocene gravels, further by the fact that the gravel is often mixed with loess /ÁDÁM-MAROSI-SZILÁRD, 1959/. The subsidence of the basin continued during the Holocene and the morphological changes of that time produced the recent picture of the basin.

DISCUSSION

To carry out the investigations, sampling was performed in three sites of the basin that represent three different formations:

1. The region of Sárszentmihány /S_1/ - the area of the paludal lime pit. Here the calcareous mud is not covered by peat.

2. The region of Nádasladány /N_1/, at the crossing of the road Nádasladány-Csór and the Nádas-canal, close to the peat excavation site.

Fig. 1. shows the sketch of the sedimentary sequence of the dug and drilled profiles at the three sites and fairly well illustrates the different states of formations. The succession process that proceeded in the basin area is best demonstrated by the section S_1, thus the biostratigraphic evaluation concerns this section.

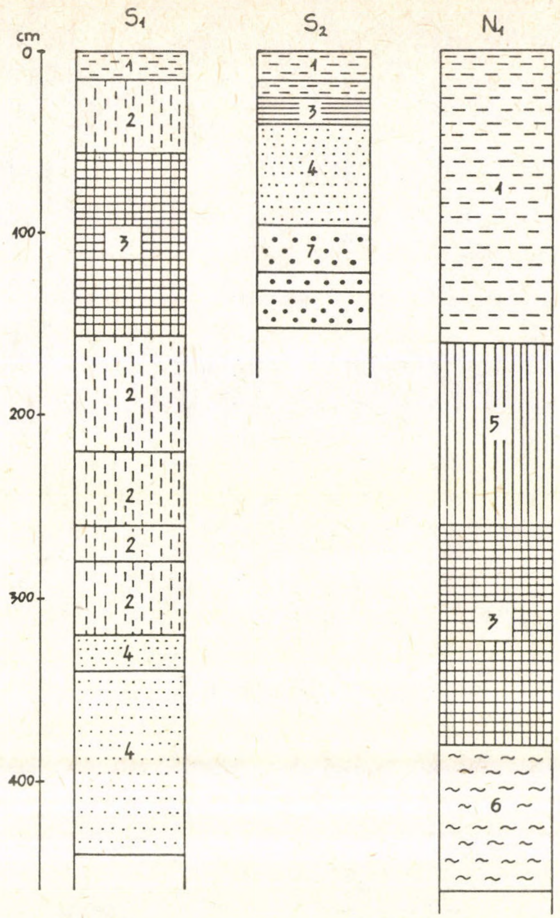

Fig. 1: Simplified sketch of the sedimentary sequence of the studied profiles. S_1 = Sárszentmihály Profile No.1, S_2 = Sárszentmihály Profile No.2, N_1 = Nádasladány Profile No 1; 1 = paludal sediment, 2 - calcerous mud, 3 = lacustrine chalk, 4 = grey sand, 5 = peat, 6 = red clay, 7 = gravel.

a. At the base of the sequence f l u v i a t i l e s a n d is found. The number of individuals in the sediments is determined by V a l v a t a p i s c i n a l i s . Regarding ecological demands, it is well-known that it favours slowly flowing to stagnant water. In addition to the predominant species, L i t h o g l i p h u s n a t i c o i d e s , which unambiguously favours flowing water also occur.

b. The sand is overlain by grey calcareous m u d intercalated locally by strips of sand. The guide species is also V a l v a t a p i s c i n a l i s but towards the top of the section its relative frequency decreases and no species indicating flowing water are found.

c. Calcareous mud is overlain by greyish-yellow l a c u s t r i n e c h a l k . The change in sediment is also reflected by faunal change the number of species increases, the former guide species is replaced by G y r a u l u s a l - b u s that lives on plant stems. In the younger section of this sediment type /closer to the surface/ the predominance of G y r a u l u s is replaced by B i t h y n i a t e n t a c u l a t a sudden increased in relative frequency.

d. The last section of the sequence reflects p a l u d a l sedimentation. The fauna is fundamentally changed: while downwards only limnic species are found, here terrestrial species also occur. This is explained by the fact that the lake reached the state when the former open water surface disappeared and became dissected by land spots. A paludal fauna proper developed with V a l v a t a c r i s t a t a as guide species. /Fig. 2 fairly well demonstrates these stages through the change of relative frequencies of marker species./

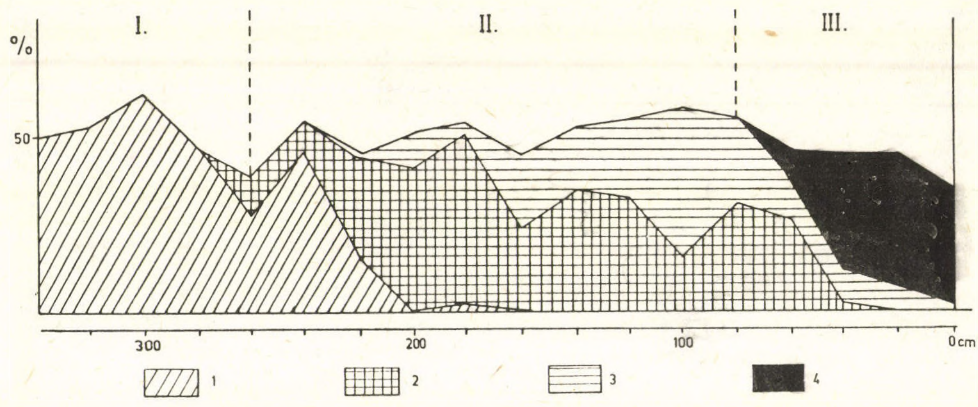

Fig. 2: Three phases of lake succession based on mollusc fauna
1 = Valvata piscinalis; 2 = Gyraulus albus; 3 = Bithynia tentaculata
4 = Valvata cristata

From the reconstructed succession sequence, regarded ideal here, the conclusion can be drawn that the profile refers to the central part of the former lake. This statement is supported by the paleobotanical evaluation of I. SKOFLEK /in: FÜKÖH, 1976/. In the calcareous mud Chara sp. nodes are found in great number, subordinately only seeds with hurling apparatus /Centaurea sp./ occur.

The results of profile analysis given above are sufficiently supported by the biostratigraphic analyses of the two other profiles. The lower strata of the S_2 section revealed a gravel cover. The fluviatile origin of the gravel is confirmed by Lithoglyphus neticoides, by the mixed fauna /partly washed together/ as well as by the foraminifers in the sediments transported from the Bakony Mountains.

Based on the thick gravel coat and on the overlying thin /14 cm/ lacustrine chalk it can be concluded that water was very shallow here along the basin margins.

In the Nádasladány region a sedimentation type totally contrasting to those above is found. In the lower part of the N_1 section, between 480 and 380 cm, red-colored calcareous clay is found. Its origin is not yet clear. From the presence of Lithoglyphus naticoides in the sediments, fluviatile origin is assured. This is supported by the fact that species with different ecological demands are also present. According to KORMOS /1909/, the age of the sediment is Pleistocene.

The fauna of the 120 cm thick calcareous mud overlying the clay is equivalent to the same sediment of the S_1 section. The same can be said on the basis of investigation of the paludal fauna. The fauna of the peat /100 cm/ between the calcareous mud and paludal formation is rather poor and can be traced back to pH changes during peat formation.

GEOMORPHOLOGICAL CHANGES AND SUCCESSION PROCESSES

The evaluation of the profiles suggests that the area, subsided in the Upper Pleistocene, began to be filled up by fluviatile sediments. This sedimentary complex is found in all the three profiles. The lower part of the sequence consists of rounded, well-sorted gravel. In the localities S_1 and S_2 this sediment is overlain by fine-grained sand. The occurrence of fluviatile sediments in different depths refers to a repeated subsidence at the Pleistocene-Holocene boundary.

The uneven topography was responsible for the differences in sedimentation at the three sites. It can be assumed that the lake-shore could exist where the gravel is found close to the surface /S_2/ and the second phases of subsidence did not affect this area. This is also reflected by the very thin /14 cm/ calcareous mud.

The profile at Nádasladány allows to conclude that due to certain reasons the riparian section was cut off/even the fluviatile sediments are different!/ thus conditions favouring peat formation came about. The second subsidence phase also affected this area, as proved by the relatively thick layers of calcareous mud.

The lake that developed to the Early Holocene was gradually filled up in living state. This is proved by the vertically oriented plant remnants in the sediments and also by the aquatic mollusc fauna. The large quantity of *Chara* nodes points to the organic origin of calcareous mud and lacustrine chalk.

The malacological material of the sediments allows us to reconstruct the succession process. Based on biostratigraphic evidence the filling of the lake can be divided into three main phases /FÜKÖH, 1977/.

Phase I.

Valvata piscinalis is the guide species. It is sensitive to limpidity, contaminations of the smallest measure may lead to its extinction. It favours the slowly flowing or stagnant water with sandy bottom and free of higer vegetation. In this phase lake water had a very high $CaCO_3$ content. One of the reasons for this /STEFANOVITS, 1956/ could be the fact that the karst spring that nowadays issue in the Csór environs flowed into this lake.

In accordance with the chemical analyses /KORMOS, T. 1909/, the $CaCO_3$ content in the calcareous mud amounts to 89.33 %, i.e. the sediments closing the first phase consist nearly solely carbonic chalk. The analysis of the organic matter carried out in the Hungarian Geological Survey /1975/ revealed only 2.1 % carbon.

For reconstruction, the presence of the species *Lymnaea ovata* is also significant, since being an euryecological species it endured the extremely calcic water, but it is very sensitive to drying and therefore, it cannot be found in the sediments close to the surface.

It can be concluded that in the area a large, relatively deep and highly calcic lake without higher plants could exist.

Phase II.

In this phase the ecological conditions considerably changed. Due to rapid sedimentation /the chemical features, see above/ water depth was gradually reduced and higher plants were established. This is indicated by the occurrence of gastropod species /G y r a u l u s a l b u s , B i t h y n i a t e n t a c u l a t a/ that live on stems. This phenomenon can be observed in the sediments of both at Sárszentmihály and at Nádasladány.

Phase III.

The lake is completely filled and this was promoted by the large-scale seasonal devastation of vegetation. This process proceeded in different manner in the three localities. In the case of the S_1 profile this involves a very short period /the thickness of the paludal sediments is 15 cm/, the N_1 profiles shows a very slow process. In the swamp sediments in addition to the typical paludal elements /L y m n a e a and P l a n o r b i d a e species/ the terrestrial species referring to drying land also occur: V a l l o n i a c o s t a t a , P u p i l l a m u s c o r u m etc.

DATING THE PHASES OF FILLING

No absolute data are available for the phases, thus only indirect geomorphological and faunistic data can be relied on.

The recent face of the Sárrét depression was developed by the Early Holocene tectonic movements. The lake period and the beginning of sedimentation followed these movements.

The phases of filling by calcareous mud and peat are be discussed below. The dating of individual phases is impossible but the overall chronology of events seems to be possible.

The starting point, as mentioned earlier, is known and the age of the last phase, i.e. the formation of the paludal sediments can be limited by means of the mollusc fauna. In the paludal sediments the number of B i t h y -

n i a t e n t a c u l a t a specimenst is gradually restricted and is replaced by the other species of the genus, i.e. by B. LEACHI (Fig. 2). The species BLEACHI is rather frequent in the Pleistocene sediments of Hungary but its relative frequency shows gradual decrease parallel with climatic amelioration. Recently it is rather rare in the Hungarian fauna.

G y r a u l u s r i p a r i u s was found solely in the plaudal sediments, this species being absent recently in Hungary. It is spread in Northeastern Europe and the southern boundary of its distribution is the Csallóköz, Czechoslovakia). Since these two species requiring a climate colder than presently occur only in this sediment típe, it can be presumed that the marsh formation followed in the first cooling (Subboreal) subsequent to the warm Atlantic period, which witnessed the formation of lacustrine chalk (?). Similar tendency can be observed in the fauna found in the course of preliminary investigation of the sediment samples of the Balaton environs (at Balatonederics).

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PALEOGEOGRAPHY OF THE POLISH CARPATHIANS DURING THE VISTULIAN AND HOLOCENE

L. Starkel

ABSTRACT

The paleogeographic changes in the Polish Carpathians during the last cold stage and the Holocene distinctly reflect the sequence of climatic fluctuations of various order. The great diversity of changes is connected with vertical zonation and mesoclimatic differentiation on the one hand and with lithology and tectonic uplift or subsidence tendencies on the other. The last cold stage, due to the cold climate and retreat of dense vegetation, was a period of degradation of elevations and deposition in depressions. On the contrary, during the Holocene most of the Carpathian slopes were forested and stable /locally sculptured by landslides/, and only valley floors were adapted to the hydrologic regime of the temperate zone. The various kinds of land use completely changed the equilibrium of natural geoecosystems.

INTRODUCTION

During the last 30 years the knowledge of the Carpathian stratigraphy and paleogeography of the last cold stage /KLIMASZEWSKI, 1961, 1967; STARKEL, 1964, 1968, 1977a, 1980; ŚRODOŃ, 1952a, 1968, 1972/, and that of the Holocene /STARKEL, 1960a; RALSKA-JASIEWICZOWA and STARKEL, 1975, STARKEL, 1977/ has been summarized. These papers were preoccupied either with stratigraphy or with paleogeography. The last decade provided a lot of new information. Especially the number of radiocarbon datings has been increasing since the launching of the new IGCP project No 158 in 1977. Numerous new peatbogs, loess and fluvial sites were studied by various methods /Fig. 1/.

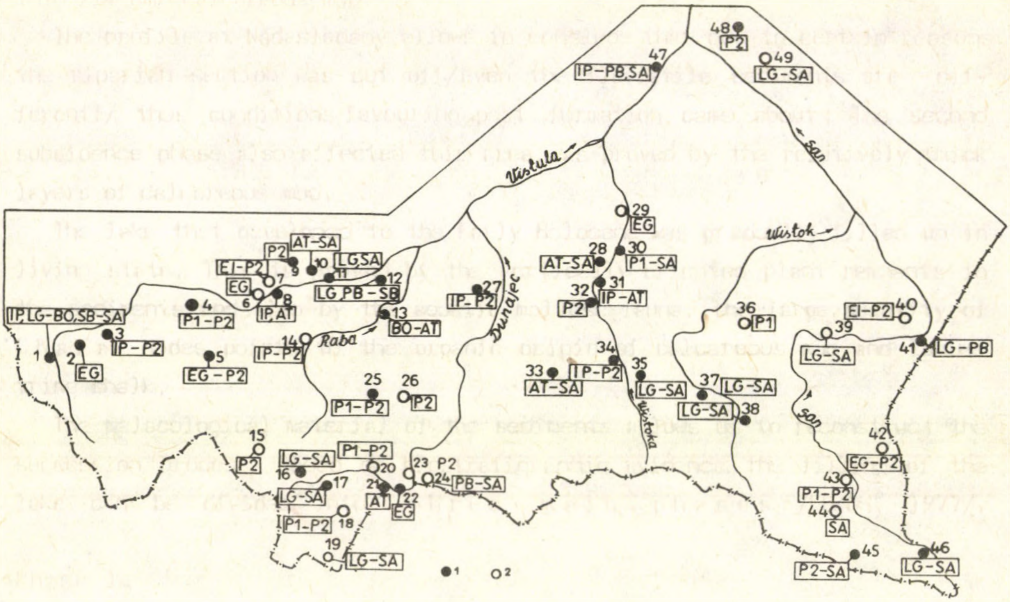

Fig. 1: Localities with described sediments of the Vistulian and Holocene /mentioned in text/, 1 = Sites dated by radiocarbon methods, 2 = Other sites /mainly dated by palynological methods/. Abbreviations: EI = Eemian Interglacial, EG = Early Glacial, P1 = Older Pleniglacial, IP = Interpleniglacial, P2 = Younger Pleniglacial, LG = Late Glacial, PB = Preboreal, BO = Boreal, AT = Atlantic, SB = Subboreal, SA = Subatlantic. Sites: 1. Drogomyśl /NIEDZIAŁKOWSKA et al. 1985/; 2. Chybie /NIEDZIAŁKOWSKA et al. 1985/; 3. Kaniów /GILOT et al. 1982/; 4. Zator /KOPEROWA-ŚRODOŃ, 1965/; 5. Wadowice /SOBOLEWSKA et al. 1964/; 6. Sciejowice /DYAKOWSKA, 1939/; 7. Zwierzyniec - 2 sites /KOZŁOWSKI, 1974/; 8. Ludwinów /SOKOŁOWSKI-WASYLIKOWA, 1984/; 9. Nowa Huta /MAMAKOWA-ŚRODOŃ, 1977/; 10. Pleszów /WASYLIKOWA et al. 1985/; 11. Branice /KALICKI-STARKEL, 1987/; 12. Grobla Forest /GĘBICA-STARKEL, 1987/; 13 Moszczenica /ALEXANDROWICZ-GERLACH, 1983/; 14. Myślenice /STARKEL, 1969/; 15. Orawka /ŚRODOŃ, 1968/; 16. na Grelu /KOPEROWA, 1962/;

17. Puścizna /OBIDOWICZ, 1985/; 18. Białka Tatrzańska /SOBOLEWSKA-ŚRODOŃ, 1961/; 19. Przedni Staw /WICIK, 1984/; 20. Brzeziny /BIRKENMAJER-ŚRODOŃ, 1960/; 21. Harcygrunt /ALEXANDROWICZ, 1984/; 22. Kąty /MAMAKOWA et al. 1975/; 23. Krościenko /KLIMASZEWSKI et al. 1950/; 24. Bryjarka /PAWLIKOWA, 1965/; 25. Dobra /KLIMASZEWSKI, 1971; ŚRODOŃ, 1968/; 26. LIPOWE /STARKEL, 1969/; 27. Szujec /SOKOŁOWSKI, 1981/; 28. Grabiny /STARKEL et al, 1981/; 29. Rzochów /LASKOWSKA-WYSOCZAŃSKA - NIKLEWSKI, 1969/; 30. Brzeźnica /MAMAKOWA-STARKEL, 1974/; 31. Podgrodzie /STARKEL et al. 1981/; 32. Pilzno /PAZDUR, 1985/; 33. Szymbark - 2 sites /GIL et al. 1974,DAUKSZA et al. 1982/; 34. Jasło-Bryły /ALEXANDROWICZ et al. 1985/; 35. Roztoki and Tarnowiec /WÓJCIK, 1987/; 36. Niebylec /BUTRYM-GERLACH 1985/; 37. Kępa /GERLACH et al. 1972/; 38. Besko /KOPEROWA, 1970/; 39. Podbukowina /MAMAKOWA, 1962; STARKEL, 1960/; 40. Orzechowice /MARUSZCZAK et al. 1972/; 41. Łodynka /HENKIEL, 1966/; 43. Zabrodzie-Solina /DZIEWAŃSKI-STARKEL, 1967/; 44. Bukowiec /unpubl./; 45. Smerek /RALSKA-JASIEWICZOWA, 1980/; 46. Tarnawa /RALSKA-JASIEWICZOWA, 1980/; 47. Piaseczno and Machów /MYCIELSKA-DOWGIAŁŁO, 1977/; 48. Łążek /MAMAKOWA, 1968/; 49. Imielty Ług /MAMAKOWA, 1962/

The northern slopes of the Carpathians, rising from ca 200 to 800-1300 m a.s.l. /only the Tatra range above 2500 m/, are exposed to the NW winds bringing rainfall, and in the past they were subjected to the influence of an extensive ice sheet from the north. During its maximum extent, the ice front was only at a distance of 300-400 from the Carpathians. Temperature and humidity in the Carpathians have a permanent vertical zonation affecting all hydrological, geomorphical, and biological processes, and geoecological boundaries shift accordingly /HESS, 1965; KLIMASZEWSKI, 1967; KOTARBA and STARKEL, 1972/. The present-day tree line runs at about 1550 m a.s.l. and the calculated snow line at 2200 m a.s.l. The W-E strike of the Carpathians leads to the frequent change of the oceanic and "dry" continental air masses. On the contrary, the gap of the Low Beskid in the main Carpathian chain facilitates the tranfer of arctic as well as subtropical sir masses in the N-S direction.

The extreme variety of meso- and microclimates combined with slope exposure related to radiation and wind directions and the inversion of temperature in the valleys should always be considered /HESS, 1965, OBRĘBSKA-STARKEL, 1970/. All these factors played a much greater role under periglacial conditions, when both temperature and humidity were factors limiting the development of plant communities. This diversity was probably similar to the present-day variability of habitats in the Khangai Mts /STARKEL and KOWALKOWSKI, 1980/. Therefore, the mountains, although located next to the ice sheet and covered with permafrost /cf. KLIMASZEWSKI, 1948/, offered a chance for the survival of numerous trees and animals in their refuges and facilitated a relatively rapid migration of species along the mountain margin and across transversal valleys and depressions.

The Polish Carpathians are built mainly of flysch rocks of various resistance, very susceptible to weathering and gravitational processes. Therefore, even under the conditions of a continuous tectonic uplift slope degradation during cool and humid phases was so strong that aggradation took place on valley floors. The main features of the Carpathian slopes with solifluction-wash glacis or even cryopediments were inherited from the last pleniglacial /STARKEL, 1969; KLIMASZEWSKI, 1971/. During the phases of reduced discharge /Holocene and interpleniglacial warmings/ rivers compensated for surplus energy by downcutting in the bedrock, and depositing in the foreland. Climatic variations generally prevented the accumulation of a continuous sequence for the Vistulian and Holocene, with the exception of higher terraces with eolian or slope sediments.

EEMIAN INTERGLACIAL

The knowledge on the last interglacial in the Polish Carpathians is very poor, due to intensive slope degradation and fluvial erosion during the last cold stage. It is assumed that some basic gravels from early-glacial localities are of that age /LASKOWSKA-WYSOCZAŃSKA, 1971; STARKEL, 1980/. The top of the Saalian slope deposits show traces of an intensive leaching and decalcification under temperate climate /DZIEWAŃSKI and STARKEL, 1967/. In the loess profiles of Pikulice, Orzechowice and Zwierzyniec at the Carpathian foreland there occur well-developed brown-earth and lessivé soils /MALICKI, 1967, MARUSZCZAK et al. 1972, CHMIELEWSKI et al. 1977/.

EARLY GLACIAL /115-75 ka BP/

No traces of early glacial cold are known from the Polish Carpathians, except loess layers with interstadial humus soils with gleying from Orzechowice /MARUSZCZAK et al. 1972/ and Zwierzyniec /CHMIELEWSKI et al. 1977/. The early glacial was a phase of fluvial activity. On the top of gravels and sands of a separate alluvial fill /Fig. 2/ there have been preserved the oxbow-lake sediments with peat of the Brørup interstadial. In Wadowice, about 6 m above the present channel, the peat layer dated at >40 ka BP contains pollen of two Picea species /dominantly Picea omoricoides/, Pinus, Alnus and also Populus, Quercus and Tilia /SOBOLEWSKA et al. 1964/. At a higher elevation /ca 500 m a.s.l./ and 26 m above the river channel in the Dunajec river valley at Kąty the peat dated at >48,200 BP contains the pollen of Picea, Abies and some admixture of Alnus and Carpinus /MAMAKOWA et al. 1975/. Of that age are the peat from Ściejowice in the loess terrace of the Vistula /DYAKOWSKA, 1939/, from Rzochów in the Wisłoka river valley /LASKOWSKA-WYSOCZAŃSKA - NIKLEWSKI, 1969/ and from the base of the Chybie alluvial fan, Oświęcim Basin with warm indicators: Abies, Carpinus and Alnus /NIEDZIAŁKOWSKA et al. 1985/. The mixed forest communities in lower elevations indicate that July temperatures reached 16-17^{0}C /probably only 1^{0} less than at present/, and the forest reached at least to 1200-1300 m a.s.l. /Fig. 3/.

OLDER PLENIGLACIAL /ca 75-50 ka BP/

This distinct cooling is represented by deluvial, soliflunctional, loess and fluvial deposits. The sequence of changes at Wadowice or Zabrodzie is typical; with the retreat of a dense vegetation cover slope wash began to be replaced by thick solifluction connected with growing permafrost /SOBOLEWSKA et al. 1964, DZIEWAŃSKI-STARKEL, 1967/. Alluvial deposition was probably synchronous. At the bottom of interpleniglacial oxbow-lake sediments /36-48 ka BP/ there are gravel beds at Brzeźnica /STARKEL et al. 1981/. The loess deposits near Przemyśl were dated by the TL method at between 80±16 and 55±12 ka BP /MARUSZCZAK, 1980; MOJSKI, 1985/. A similar TL age is shown by some loess-like loams on the high terraces in the valleys /BUTRYM-GERLACH, 1985; ZUCHIEWICZ, 1985/. We may expect a penultimate

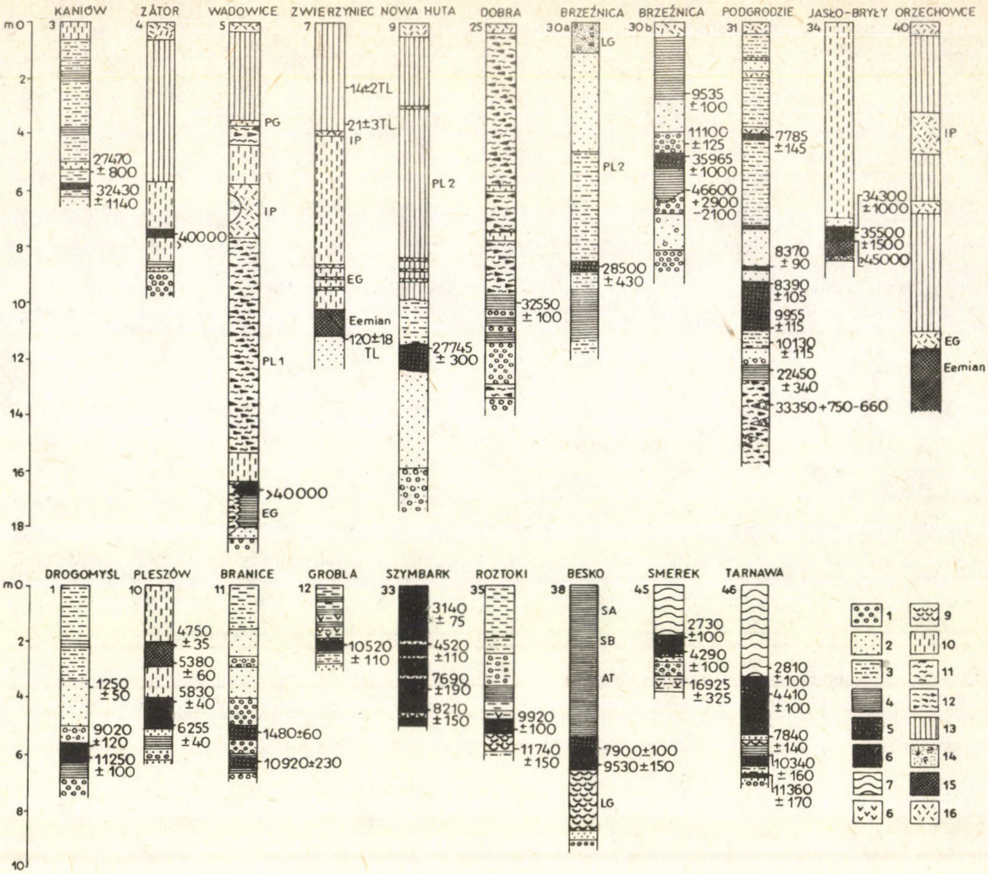

Fig. 2: Selected profiles of Vistulian and Holocene sediments in the Polish Carpathians.

1 = gravels, 2 = sands, 3 = sandy alluvial silts, 4 = clays, 5 = peat muds, 6 = peat, 7 = Sphagnum peat, 8 = gyttja, 9 = lacustrine chalk, 10 = loamy deluvia, 11 = solifluction loam, 12 = solifluction loam with debris, 13 = loess, 14 = eolian sand, 15 = forest soil, 16 = arctic type soil. Age in radiocarbon years. For abbreviations see Fig. 1.

glaciation in the Tatra Mts., which was probably reflected in the deposition of gravels underlying the interpleniglacial boreal peat at Brzeziny /BIRKENMAJER-ŚRODOŃ, 1960/, and in the lower gravel horizon of the glacifluvial fan at Zakopane /GUZIK-JACZYNOWSKA, 1959/.

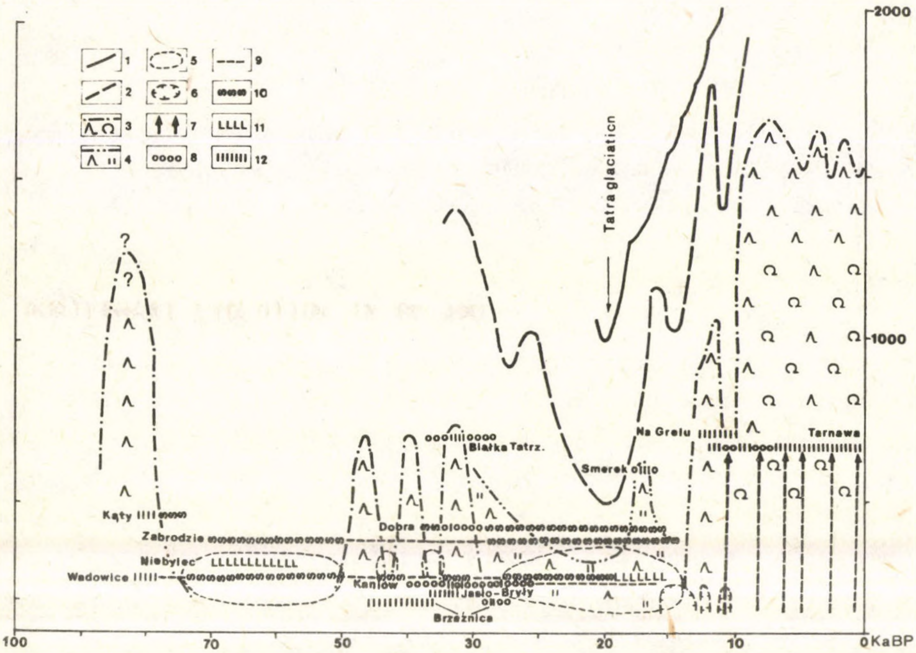

Fig. 3: Changes of vertical zonation in the Polish Carpathians during the Vistulian and the Holocene

1 = extension of local Tatra glaciers /after KLIMASZEWSKI, 1967/; 2 = limit between cryonival and alpine tundra zones; 3 = upper timber line /dominance of boreal and decidous trees shown by symbols/; 4 = hypothetic tundra-open woodland-steppe zone; 5 = zone of loess deposition; 6 = zone of dune formation; 7 = phases of increased fluvial activity; 8-12 = deposits at localities shown in Fig. 1: 8 = alluvia; 9 = deluvia; 10 = solifluction deposits; 11 = loess; 12 = organic deposits

Two only localities with a Dryas flora show a typical spectrum of a treeless vegetation: at Zator elevated 240 m a.s.l. /KOPEROWA-ŚRODOŃ, 1965/ and in Ludwinów at 200 m a.s.l. /SOKOŁOWSKI-WASYLIKOWA, 1984/. Both of them indicate that summer temperatures were not higher than +10°C.

INTERPLENIGLACIAL PERIOD /59-29 ka BP/

The number of localities from that period reaches twenty, but due to the limitation of the radiocarbon method it is difficult to distinguish various deposits older than the Denekamp interstadial. The dated horizons are mainly oxbow-lake deposits. During the older warming /Hengelo ?/ in the Wisłoka river valley at Brzeznica /48-36 ka BP/ there dominated an open forest with 50-60 % AP with Pinus, Pinus cembra and Larix /MAMAKOWA-STARKEL, 1974/. A similar forest is described from the locality Jasło-Bryły elevated 220 m a.s.l. dated at 35,500±1500 BP /ALEXANDROWICZ et al. 1985/. From the Denekamp a fluvial member with an organic layer is described from Dobra at the elevation of 420 m with a rich forest /Picea, Pinus, Larix, Pinus cembra and Alnus/ and amounting to 15-70 %, dated at 32,550±450 PB /SRODON, 1968/. From the altitude of 700 m there was described an open forest with Pinus cembra, Pinus, Larix and Betula at Białka Tatrzańska /SOBOLEWSKA-ŚRODOŃ, 1961/. The upper tree line was probably rising above 700 m. In Myślenice, at only 300 m a.s.l., the percentage of tree pollen amounts to 85 % /ŚRODOŃ, 1968/.

Due to a certain stabilisation of slope, the interpleniglacial period was in general a phase of high fluvial activity and dissection of older solifluction deposits /Fig. 2/. The alluvial fans are known from Kaniów in the Vistula river valley /32,430± 1140 BP - GILOT et al. 1982/, and from Szujec in the Dunajec river valley /31,425±530 - SOKOŁOWSKI, 1981/. Dissections were registered at Brzeźnica in the Wisłoka river valley /MAMAKOWA-STARKEL, 1974/ and at Tarnobrzeg in the Vistula valley /MYCIELSKA-DOWGIAŁŁO, 1977/.

The existence of a very thick active layer, or even the retreat of permafrost induced landslides reported from Wadowice /SOBOLEWSKA et al. 1964/ and Myslenice /STARKEL, 1969/. Lixiviated horizons are known from many localities in the San and Strwiąż river valleys /DZIEWAŃSKI-STARKEL, 1967; HENKIEL, 1966/. The loess profiles at the foreland show the horizon of subarctic soils with Aurignacian culture /CHMIELEWSKI et al. 1977/ and with mammoth bones dated at 25-29 ka BP by the FCL/P/Coll. method /LAS-

KOWSKA-WYSOCZAŃSKA, 1971/. The climate of the interpleniglacial period was very changeable. During the warmings, July temperatures were high enough for forest growth /up to 13-15°C along the margin of the Carpathian Foothills.

YOUNGER PLENIGLACIAL /29-13 ka BP/

The last cool phase with a maximum extent of the Scandinavian ice sheet was not only the coldest but also the driest one /MARUSZCZAK, 1980, VELICHKO, 1984/. But this does not mean that the climate was continually cold, there exist many indications of its fluctuations.

The first cool wave started after 29 ka BP and the typical tundra spectra in the Subcarpathian basins are known from Nowa Huta /27,745±300 BP - MAMAKOWA-ŚRODOŃ, 1977/ and from Brzeźnica /28,500±430 BP - STARKEL,1981/. In the Wisłoka valley this type of a treeless vagetation is dated also at 24-25 ka BP /PAZDUR, 1985/. But the spectra from Podgrodzie at the margin of the Carpathians dated 22,450±340 BP show a high percent of AP, up to 40-50 % /STARKEL et al. 1981/. Below this horizon there was found the wood of Picea vel Larix dated by M.A. Geyh in Hannover at 26,980±345 BP. Therefore, the old conception of local refuges mentioned also by ŚRODOŃ /1968/ and STARKEL /1980/ seems to be very realistic. Under favourable mesoclimatic conditions boreal trees could survive as small open woodlands. This view may be supported by the locality Łążek on the northern margin of the Sandomierz Basin, where 25,580±3270 BP there existed an open forest with Betula, Larix and Alnus /MAMAKOWA, 1968/, and also by the presence of pine or spruce charcoals at the mammoth hunters site at Zwierzyniec in Cracow dated at 23-20 ka BP /KOZŁOWSKI et al. 1964/. In higher elevations there dominated the alpine tundra /ŚRODOŃ, 1968/ but, unfortunately, we have no radiocarbon datings from those sites.

The slope deposits in the Carpathians show a very distinct vertical zonation. Above 700-1000 m a.s.l. frost weathering was dominant and we find many records on the thick debris covers, frost riven cliffs and cryoplanation terraces /KLIMASZEWSKI, 1948, 1971; BAUMGART-KOTARBA, 1974; STARKEL, 1969, 1980/. This belt reached the snow line at ca 1500 m, and the tongues of the Tatra glaciers, which moved down to 900-1150 m a.s.l. /KLIMASZEWSKI, 1967/. At the elevation of 300-700 m a dominant deposit of that period is solifluction loam with coarse debris, sometimes 10 m thick, overlain with slope wash deposits. This vertical belt may be called the cryohumid zone, due to the dominance of solifluction processes. In some localities /Krościenko, Dobra/ a sequence of at least four thicker debris horizons, separated by thin clayey layers was formed. It is connected with a colder climate with a thinner active layer /KLIMASZEWSKI, 1971/. At the close of the pleniglacial there was a distinct tendency towards a drier climate reflected in the rising of the cryoarid vertical zone, and the formation of the loess top layer in Wadowice proved by a different composition of heavy minerals compared with the solifluction layers below /SOBOLEWSKA et al. 1964/. The slope wash dominated in the Carpathian foothills /KRYGOWSKI, 1963/ as well as in the intramontaneous depressions /CEGŁA, 1964/. The prevailing drier climate is also shown by an upper solifluctional layer rooted in the level of slope hooks in the middle part of the slope in Solina /STARKEL, 1965/.

The lowest vertical belt at 200-350 m may be called cryoarid with the dominance of loess deposition. The composition of heavy minerals /RACINOWSKI, 1976/ indicate the dominance of short distance transport. The loess deposition continued from 28 to ca 15-12 ka BP with a deposition rate 0,5-1 mm per yr /MARUSZCZAK, 1985/. The younger TL datings from Zwierzyniec give the age 14±2 ka BP /MOJSKI, 1985/. In the Jasło-Sanok Depression, much like in the whole transversal depression on the Low Beskid, there is a lot of evidence of a strong wind activity in the form of N-S directed close deflational depressions, some hundreds of metres long and filled with late glacial lacustrine sediments /GERLACH et al. 1972/ and ventifacts on hills /GERLACH et al. 1983/.

The fluvial deposits in the upper river courses show a distinct interdigitation with slope solifluction deposits /STARKEL, 1960b, 1965/. But in the main valley, especially in the foreland, the lack of channel deposits indicates a change in stream activity and a tendency to downcutting. The last dated fluvial deposits of the Vistulian terrace are 28-27 ka old and show a quick sandy-silty deposition of snowmelt floods /MAMAKOWA-STARKEL, 1974/. In the northern part of the Sandomierz Basin a younger fill covered with Holocene alluvia is represented by the sands of the braided river /MYCIELSKA-DOWGIAŁŁO, 1977/. The sandy muds at Podgrodzie dated at 22,450±340 BP also show a relatively low position of the river channel at that time /STARKEL et al. 1981/. Probably two factors were acting simultaneously: the lowering of the Vistula base level with a retreating ice sheet, and increasing continentality. The Grel peatbog along the Czarny Dunajec river in the Nowy Targ Basin, dated back to the Bølling, lies on the erosional bench cut in the fluvioglacial fan /KLIMASZEWSKI, 1961/.

The preservation of tree patches on the northern margin of the Carpathians indicate that summer temperatures attained locally 10^{0}C. During warmings the temperature was probably higher, as it is shown by the organic horizon in the alluvial fan at Smerek in the Bieszczady Mts. dated at 16,925±325 BP /RALSKA-JASIEWICZOWA, 1980/. The pollen spectra with 60-80 % AP and the presence of Pinus cembra, Larix, Betula and Alnus viridis indicate that the upper tree line was located at 600 m a.s.l. This means that on the Carpathian margin, 400 meters below, July temperatures reached ca 13^{0}C. This reconstructed cold continental climate is in agreement with the data from the western margin of the USSR presented by GRICHUK /1982/ and VELICHKO /1984/. But indications of strong eolian activity point to an extremely great variety of the geoecosystems under cold climate with permafrost.

LATE GLACIAL /13-10 ka BP/

The warming of climate and the melting of permafrost caused changes in the vegetation and in geomorphic processes. The spreading of boreal forest with Pinus silvestris, Pinus cembra, Larix, Picea and Betula is pronounced in the organic deposits of the Allerød /MAMAKOWA, 1982; KOPEROWA, 1962; RALSKA-JASIEWICZOWA, 1980/. The forest reached the elevation of ca 1000 m a.s.l., but on steeper slopes the rewashed block fields delayed forest invasion /STARKEL, 1960b/. The composition of the forest indicates that these species were preserved in local refuges.

Slope wash and solifluction were stopped by dense vegetation and the delivery of new debris to river channels replaced by deep landslides favoured by the deep circulation of water /STARKEL, 1960a; PAWLIKOWA, 1965; ŚRODOŃ, 1952b/. Therefore, the rivers changed their mechanism from braided to meandering. These large meanders with fills dated between 12 and 10 ka BP were investigated in the San /SZUMAŃSKI, 1982/, Wisłoka /STARKEL et al. 1981/ and Vistula river valleys. In the Vistula valley as well as in the middle course of the San valley /of. MAMAKOWA, 1962/ the pre-Allerød erosion reached even below the present river level. In the Wisłoka valley, on the margin of the Carpathians, the Late glacial erosional plain formed by large-scale meanders is 4-6 m above the present mean river level /STARKEL et al. 1981/, which may be explained as being the result of a young tectonic uplift. Similar erosional benches rising above the Allerød channel east of Cracow in the Vistula valley probably indicate the tendency to aggradation and braiding during the Younger Dryas /KALICKI-STARKEL, 1987/ when, due to cooling, the upper tree line fell to ca 600-700 m a.s.l. /ŚRODOŃ, 1972; RALSKA-JASIEWICZOWA, 1980/.

The rise in precipitation produced lakes in the deflational depressions of the Jasło-Sanok Basin, where calcareous gyttja and lacustrine chalk were deposited /GERLACH et al. 1972; KOPEROWA, 1970; WÓJCIK, 1987/. This means an increase in the leaching of soils. In the Subcarpathian Basins, during cooler phases of the Late glacial, dunes were formed up to 25 m high /WOJTANOWICZ, 1968/. These dunes accompany the river channel, which points to braided rivers as the main sources of material. Only at Jęzor in the Oświęcim Basin two distinct phases of eolian activity are divided by peat dated at 11,740±100 a BP /STARKEL, 1980/. In the east, at Imielty Ług, high sand content in organic layers indicates that the dune phase lasted up to the end of the Preboreal.

HOLOCENE /10-0 ka BP/

The rapid warming and expansion of a dense vegetation is visible not only in the pollen diagrams ca 10 ka BP /RALSKA-JASIEWICZOWA, 1980, 1983/ but also in the stabilisation of river channels. The last coarser sandy deposits were dated between 10,375±125 a BP and ca 9900 a BP /STARKEL, 1977a/. Rapid warming is also reflected in the Tatra Mts at 1668 m a.s.l. where 9900±120 BP mineral clay gyttja was replaced by detritus gyttja /WICIK, 1984/. The oldest small-size paleomeanders in the San river valley were dated at the bottom of the fill at 8560±100 a BP /SZUMAŃSKI, 1985/. The warming continued, and ca 9300 BP, it reached the first peak reflected in the oxygen curve from the calcareous tuff in the Cracow Upland /A. PAZDUR et al. 1987/. The forest spectrum was still typical of the taiga, which indicates a continental climate with cold winters and low precipitations. Therefore, the deciduous trees, such as Ulmus, Corylus and Quercus, penetrated very slowly from the far refuges following the Carpathian margin or crossing transversal gaps in the Carpathian chain /RALSKA-JASIEWICZOWA, 1983/.

Ca 8500-8400 a BP there followed a rapid change of climate reflected in the deposition of alluvial loams or sandy fans over the peatbogs or oxbow deposits at Podgrodzie /MAMAKOWA-STARKEL, 1974/, Tarnawa /RALSKA-JASIEWICZOWA - STARKEL, 1975/, Roztoki /WÓJCIK, 1987/ and many other localities. This turn coincides with the spreading of dense deciduous forests and downcutting of main river channels /STARKEL et al. 1981/. The phase of high frequency floods continued probably till 7700 BP and is synchronous with the glacial advance of the Venediger stage in the Alps /STARKEL,1985/. During that phase numerous landslides were formed, among others at Szymbark, dated before 8210±150 a BP /GIL et al. 1974/ and in Harcygrund at 7750±130 a BP /ALEXANDROWICZ, 1984/.

The Atlantic phase had a relatively stable warm climate which is visible in the dominance of deciduous forests and the formation of vertical zones. The former ca 300 m higher position of the upper forest limit to ca 1850 m a.s.l. is indicated by the existence of the oval karren, later on dissected after the retreat of forest /KOTARBA-STARKEL, 1972/. From that period there are the wellknown calcareous tuffs in the Czechoslovak Carpathians /LOŽEK, 1975/, as well as the lacustrine chalk /ALEXANDROWICZ-GERLACH, 1983/. The stable fluvial regime shows only one phase with a higher flood frequency ca 6500-6000 a BP manifest in the buried oaks in the Wisłoka valley /STARKEL et al. 1981/, and the abandoned channel of the Vistula at Pleszów /WASYLIKOWA et al. 1985/.

The next distinct change with a cooling and a rise in precipitation followed at the advent of the Subboreal period ca 5000-4400 a BP. It is marked by a rise of groundwater table and a change of vegetation /RALSKA-JASIEWICZOWA - STARKEL, 1975/. The relict deciduous forest with Tilia has been preserved at 500-600 m a.s.l. in the mesoclimatic warm belt 100-200 m over the valley bottom of the Poprad river /OBRĘBSKA-STARKEL, 1967/. In the Nowy Targ Basin, at 600 m altitude the phase with an expansion of Picea is recorded between 4900 and 3700 a BP /OBIDOWICZ, 1985/. It was followed by the phase with Carpinus, Abies and Fagus. At that time in Szymbark at 450 m a.s.l. there was also a phase of the dominance of spruce /GIL et al. 1974/. Many new landslides came about and older ones were activated /cf. GIL et al. 1974/. In the Vistula river valley, downstream Cracow, the whole paleochannel system was abandoned due to avulsion hundreds of years before the layer in the fill dated at 4380±110 a BP was deposited /GĘBICA-STARKEL, 1987/.

The Subboreal is a period of expansion of the Neolithic agriculture in the Carpathian foothills, and, therefore the reflection of climatic variations is less clear /RALSKA-JASIEWICZOWA - STARKEL, 1975/. At the beginning of the Subatlantic there are indications of an increased flood activity /STARKEL et al. 1981; KLIMEK, 1987/. In the Ropa river valley the incision phase is dated before 2675±60 a BP /DAUKSZA et al. 1982/. The overflooding and aggradation continued during the Roman period when there was a coincidence of higher precipitations and deforestation dated between 2nd century BC and 3rd century AD from the Vistula and other river valleys. Human impact is clearest at Branice, where the remains of forest clerance from 1st and 2nd century AD were found /KALICKI-STARKEL, 1987/. In many localities the peatbogs were covered with alluvial loams /1950±50 a BP at Tarnowiec - cf. WÓJCIK, 1987/.

During the last millennium deforestation increased in higher mountains, too. This is seen in the aggradation in the subsidence basin at the foreland of the Silesian Beskid /NIEDZIAŁKOWSKA et al. 1985/. In the Wisłoka valley, the channel deposits dated 900-400 a BP buried the loams of the former alluvial plain /STARKEL et al. 1981/. In Cracow, loamy flood deposits from 11th-12th centuries were found in archaeological sites /RADWAŃSKI, 1972/. Soil erosion and overbank deposition started immediately after the advent of cultivation. As the best indication there may serve the loamy fill of the small creek valley floor at Bukowiec in the upper San catchment. The organic horizon at the base of the 2 m thick loam was dated at 460±65 a BP which coincides with historical data on the colonisation of this region at the turn of the 15th-16th centuries.

During the 15th-19th centuries the picture is even more complicated. The overpopulation of the Carpathians is superimposed on the humid and cool phase of the Little Ice Age /LAMB, 1977/. Higher flood frequency caused river channel widening and deepening as well as straightening and avulsions /STARKEL, 1960a, 1983; STARKEL et al. 1981/. In 18th century the channels show a tendency to braiding /SZUMAŃSKI, 1982/, which coincides with the cultivation of potatoes. From the mid-19th century the mountain foreland was affected by river regulation, which caused the next phase of downcutting reaching in some valleys 2-3 metres /STARKEL et al. 1981; KLIMEK, 1979/. In transversal depressions the leading role among the present-day processes is played by eolian activity /GERLACH, 1976/. Any change in economy and landuse is reflected in the transformation of the alluvial plain and river channels.

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FOSSIL ROCK GLACIERS IN THE POLISH TATRA MOUNTAINS: ORIGIN AND AGE

A. Kotarba

ABSTRACT

In recent years fossil rock glaciers have been studied in the Polish Tatra Mountains. Depending upon the broad definition of these features, a large part of the high-mountain relief is classified as rock glaciers, even if there are typical areal deglaciation features. According to author's concept there are no indications suggesting the existence of large areas covered by rock glaciers. Only small fragments of the mountains were affected alpine permafrost. They are related to Younger Dryas - the latest cold period in the Last Glacial history of the Tatra Mountains.

WHAT ARE ROCK GLACIERS?

Modern alpine research on active rock glaciers in which geophysical, geodetic, photogrammetric and hydrological methods were employed shows that they are diagnostic landforms of permafrost in high mountain areas. They are useful for reconstructing climatic conditions during the Late Glacial and the Holocene of the Alps /KERSCHNER, 1978/. The definition accepted by HAEBERLI /1985/, earlier formulated by WAHRHAFTIG and COX /1959/ says that rock glaciers are "perennially frozen debris masses which creep down mountain slopes in some ways similar to the behaviour of lava streams. They are, therefore, long term and large-scale natural experiments on the creep behaviour of non-consolidated, ice-rich sediments". Some authors do not accept other than interstitial ice in "true" rock glaciers /HUGHES, 1966/.

The above definition was confirmed by detail field studies in the Austrian Alps /HAEBERLI-PATZELT, 1982/ and in Swiss Alps /BARSCH, 1977/ and give us a new tool for paleogeographical reconstructions when inactive or fossil rock glaciers are found in certain mountains, nowadays free of active features. This viewpoint is not accepted by all researchers. In the literature concerning rock glacier origin there exists a definition which includes the concept that rock glaciers do not only consist of interstitial ice, but also very often are composed of relatively clean glacier ice mantled by debris. Such forms are called i c e - c o r e d r o c k g l a - c i e r s /WHITE, 1976/. Some authors conclude that the term ice-cored moraine and rock glaciers are synonymous /BARSCH, 1971/. Whalley /1983/ accepts the concept that "terminal moraines, rock glaciers, and protalus ramparts are related forms, only differing in topography". As a consequence of this broad definition of rock glaciers WHALLEY /1983/ suggests: "it would appear that they are not good indicators of climatic conditions /../ and the presence of permafrost may help to preserve a rock glacier but does not cause it".

For other researchers the truth lies midway. They agree with the idea of r o c k g l a c i e r as debris mass cemented by interstitial ice or glacier ice mantled by debris, but in both cases it must be affected by downslope displacement. ØSTREM /1971/ says: "the slope must possibly be taken into account - a certain critical angle may exist which must be exceeded to start a downslope movement". It is also accepted by UNESCO /1970/ that a rock glacier is defined as a glacier-like mass of rocks slowly moving downslope. The movement is explained by the presence of interstitial ice or glacier ice mantled by debris and in certain cases both forms occur within the same object. This broad definition has important influence on the genetic interpretation of glacial drift deposits in mountain areas and creates serious misunderstandings. Large areas glaciated during the Pleistocene are interpreted as rock glaciers, even if they have not the typical topography produced by debris displacement. Fossil rock glaciers are easily recognized on both steep and flat surfaces.

ROCK GLACIERS IN THE POLISH TATRA MOUNTAINS

In the Polish Tatra Mountains one can observe fossil rock glaciers characterized by specific micromorphological pattern with regular furrows, and systems of ridges connected with the flow processes. Such features were recognized also on the Slovak part of the Tatras /NEMČOK, MAHR, 1974/. Large areas covered by glacial deposits, however, are free of such pattern. Most of them are located on flat, and broad valley bottoms above actual timberline. This morphological criterion makes it possible to distinguish easily areas of areal deglaciation and true rock glaciers.

In March, 1987, two papers were published on rock glaciers in the Polish Tatra Mountains by DZIERŻEK and NITYCHORUK /1986/ and DZIERŻEK, LINDNER and NITYCHORUK /1986/. These papers concentrate on Late Quaternary deglaciation of the High Tatras. The above authors accept the broad definition of the term "rock glacier". As a consequence of that viewpoint a large area of the High Tatras is classified as fossil rock glaciers and four types of them were distinguished on the basis of their location within major glacial features; v a l l e y, c i r q u e, s u b s l o p e and c o l r o c k g l a c i e r s are suggested by DZIERŻEK and NITYCHORUK /1986/. Simultaneously, the above, so-called valley and cirque rock glaciers are classified on the detail geomorphological map of the Tatra Mountains as cover moraine and recessional moraines /KLIMASZEWSKI, 1985/.

DZIERŻEK and NITYCHORUK /1986/ explain the origin of valley and cirque rock glaciers as features related to an intensified accumulation of slope deposits on ice, later affected by insignificant transport downstream, due to melting of ice body. These features are correlated with the Oldest Dryas /valley rock glaciers/ and with the Younger Dryas /cirque rock glaciers/. This new interpretation of deglaciation features in major Tatran valleys: Gąsienicowa, Pańszczyca and Pięc Stawów Polskich causes many serious doubts. The most important are the following:

1. Valley and cirque rock glaciers have microtopography of a stagnant ice landscape, with undrained depressions filled with small lakes /i.e. Dwoisty staw Lake, Kurtkowiec in Gąsienicowa valley and Czerwony Staw Lake in Pańszczyca valley/.
2. The gentle slopes of the valley bottoms as well as small thickness and spatially discontinuous glacial deposits suggest that these deposits

were not displaced downslope during deglaciation. Debris accumulation does not reach the critical thickness of 30-100 m which makes plastic deformations possible.

3. Only subslope and col rock glaciers can be termed as true rock glaciers, formed on slopes inclined more than 5^{0}, rich with debris material and affected by downslope movement. According to DZIERŻEK and NITYCHORUK /1986/ subslope forms were created during the Younger Dryas, while col rock glaciers located beneath the mountain cols at avalanche routes were formed during the Little Ice Age. The age of col rock glaciers is also questionable. This will be shown below.

Detailed studies on active rock glaciers in the Austrian Alps and Swiss Alps /HAEBERLI, 1973; HAEBERLI-PATZELT, 1982/ show that rock glaciers are creep phenomena of discontinuous alpine permafrost, even if they exist side by side with ice glaciers and are the typical features of periglacial geomorphology in dry and cold mountains. Boundary conditions for the existence of rock glaciers were established. These are: mean annual air temperature $-1^{0}C$ to $-2^{0}C$ and an annual precipitation less than 2500 mm /see Fig. 1/. Fossil rock glaciers in the Polish Tatra Mountains have developed as lobate forms on debris slopes below rockwalls and are located within an altitudinal belt between 1500 and 1900 m /KOTARBA, 1986/. In some part of the mountains they are tongue-shaped. According to the paleoclimatic reconstruction by HESS /1968/ boundary conditions which were able to generate rock glacier formation occurred during the Younger Dryas /Fig. 2/. This fact suggest s that most of the recognized true rock glaciers /subslope and col r.g. according to DZIERŻEK -NITYCHORUK/ were formed at that period. During the Holocene warming this limit was shifted to an altitude of 2200 m i.e. in the bare rockwalls and rocky summit zone. This part of the mountains does not deliver the quantity of material necessary for rock glacier formation and local topography does not favour debris accumulation. The previously mentioned suggestion, that so-called col rock glaciers were formed during the Little Ice Age cannot be confirmed in the light of boundary conditions during the whole Holocene. We have no climatic evidence to conclude that mean annual air temperature isotherm $-2^{0}C$ was lowered 200-300 m below its actual position during last 600-100 years. The comparison of dendroclimatological reconstructions of summer temperatures in the Alps and the Tatra Mountains from 1741-1965 made by BEDNARZ /1984/ suggests clear analogies in these mountain ranges, but the cooling of climate in the

High Alps documented by maximum glacier advances in the years 1805-1845 /1855/ was marked in the Tatras as an alluviation phase. A great number of big debris flows was triggered at that time /JONASSON-KOT-KOTARBA, in press/. This conclusion is based on lichenometric datings. It is quite probable that Little Ice Age was only marked in the Tatras by a little larger multiannual snow patches /glacierets/ than these existing nowadays on shadowed debris slope apexes in the High Tatra Mts /WDOWIAK, 1961/.

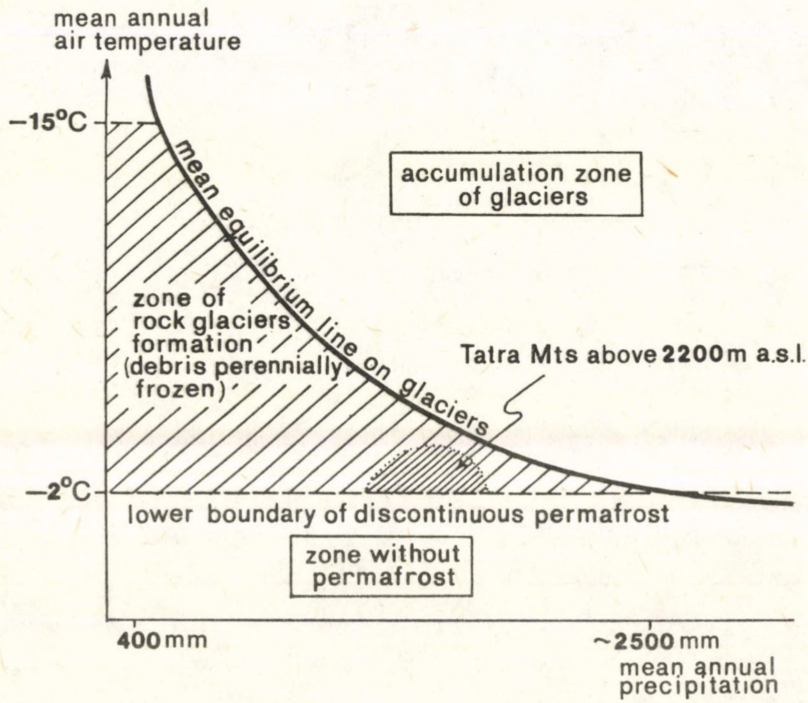

Fig. 1: Boundary conditions for rock glacier formation in the Tatra Mountains against HAEBERLI's /1985/ diagram of cryosphere structure

Fossil rock glaciers characterized by surface topography which is reminiscent of pictures of moving viscous materials and formed in the shape of distinct lobes or small tongues remain in different relation to recessional moraines. Four typical situations can be shown. They describe local conditions for rock glacier formation during and after deglaciation:

/1/ lobe-like or tongue-like features developed from steep sides of moraine ridge of the youngest recessional stage /an example in the

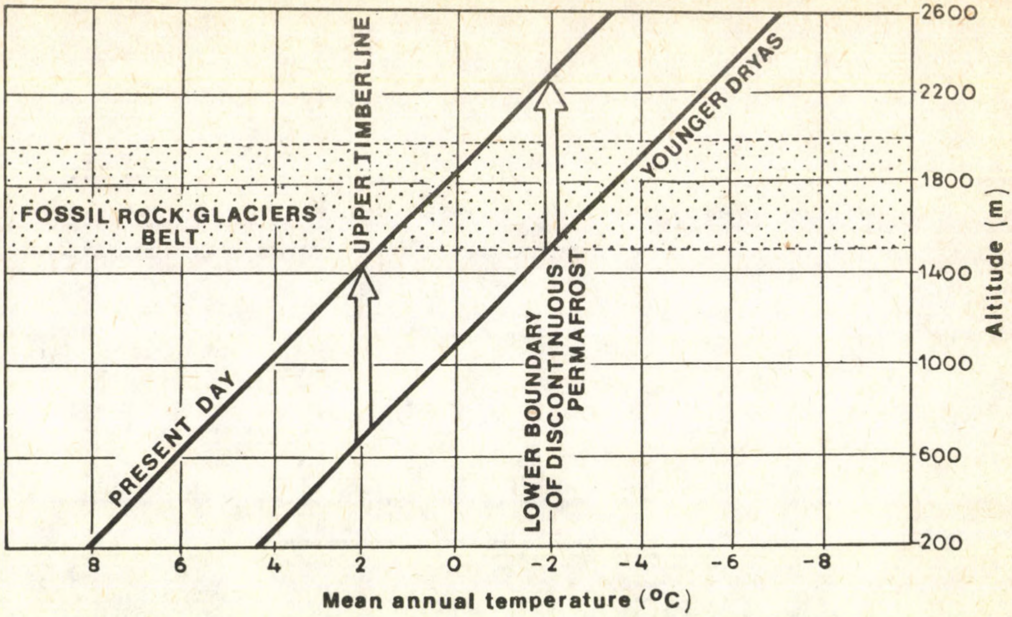

Fig. 2: Potential boundary of discontinuous permafrost in the Polish Tatra Mountains during the Younger Dryas and present-day. Mean annual temperature shifts for altitudinal belts according to HESS /1968/

Buczynowa Dolinka valley/,

/2/ lobe-like feature deposited on the top of recessional moraine ridge and generated on debris slope /an example in the Gąsienicowa valley in the nearest vicinity of Dwoisty Staw Lake/,

/3/ tongue-like rock glacier formed in small tributary valley and deposited on lateral moraine of main valley glacier /an example in the Pańszczyca valley beneath Wierch pod Fajki Mt/,

/4/ lobe-like or/and tongue-like rock glaciers formed inside the youngest recessional moraine in the uppermost part of valley heads. These features were generated from debris slopes /examples in the Western Tatra: Starorobociańska, Kościeliska, Chochołowska valleys/.

All the above situations suggest that rock glaciers were formed during the youngest stage of deglaciation. This stage should be correlated with the last cold period, at the end of the Last Glacial; Younger Dryas /Egesen stage in the Alps/.

Due to rapid change in the position of the climatic belts at the beginning of the Holocene /KOPEROWA, 1962; KRUPIŃSKI, 1984/ rock glaciers started to become inactive and fossil. Transport of debris from the rockwalls and rocky slopes across the snow patches was effective on debris slopes. As a result protalus ramparts were formed both in the High and Western Tatras. They are of the same age or younger than rock glaciers which developed during the deglaciation. Full sequence of slope units in the high-mountain part of the Tatras is shown on Fig. 3.

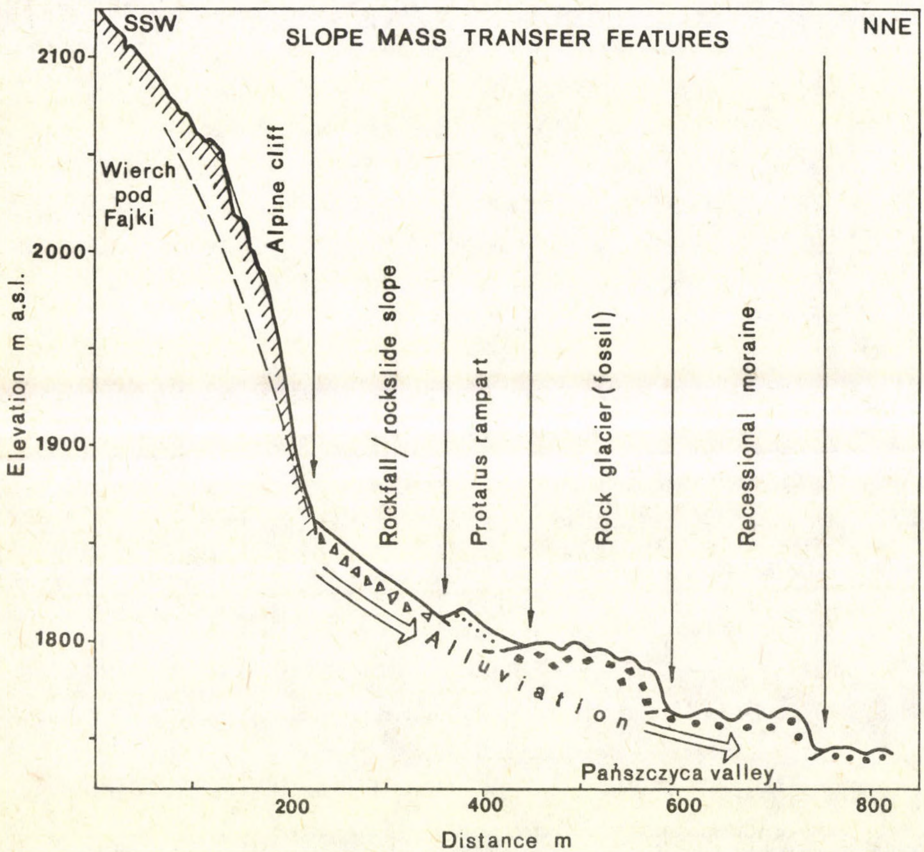

Fig. 3: Sequence of the slope mass transfer features in the High Tatra Mountains. The complex of alluvial forms /debris flow tracks/ can occur across all slope units

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HOLOCENE CLIMATIC CHANGES AS REFLECTED IN MORPHOLOGY AND ALLUVIA OF THE UPPER VISTULA VALLEY

T. Kalicki

ABSTRACT

The paper shows how climatic stages are reflected in increased river activity during the periods 8500-7500, 6500-5900, 5000-4500, 2800-2200, 1150-650 BP in morphology and in the alluvia of the Vistula valley in the vicinity of Cracow. These periods were characterized by channel changes the Vistula, formation of new alluvial series, changes in sedimentation on flood-plain, accumulation of tree trunks in alluvium or burial of cultural layers by flood-plain deposits.

INTRODUCTION

The paper aims at presenting the climatic fluctuations and the related changes in stream activity. The material is gathered in the Vistula river valley in the vicinity of Cracow. This area is investigated as part of the Problem IGCP - 158A "Paleohydrological changes in the temperate zone in the last 15,000 years" (STARKEL 1983a).

The Vistula river flows through the western part of the Sandomierz Basin after having cut its way through the Cracow Gate limestones. There are two high Vistulian terraces in the valley and a flood-plain, narrow in the Gate and wide in the Sandomierz basin. This flood-plain preserves numerous paleochannels and has a very complicated geological structure (GĘBICA, STARKEL 1987, KALICKI, STARKEL 1987, RUTKOWSKI 1987).

The changes of climate, humidity and temperature during the Holocene are reflected not only in the changes of vegetation but also in the functioning of the whole geographical environment (CHOTINSKI, STARKEL 1982). Climatic changes and, subsequent alterations in the hydrological regime of rivers are clearly visible in the morphology and the alluvia of numerous valleys in the temperate zone (STARKEL 1983b).

The cycle of fluctuation in stream acitivity is clearly visible in 2000 years. It is possible to identify stages of increased fluvial activity: 8500-7500, 6500-5900, 5000-4500, 2800-2200, 1150-650 BP (STARKEL 1986). These stages are synchronous with lake level fluctuations, mountain glacier retreats, enlivened landslides (STARKEL 1983a, b). The recovery of the system of the river channels took peaces in these periods, related to cutting single meanders (TRAFAS 1975) or avulsion (MIKE 1975, BORSY, FÉLEGYHÁZI 1983, STARKEL, KALICKI 1984, GĘBICA, STARKEL 1987). These periods were also characterized by intensified erosion and accumulation of new alluvial series within the floodplain (ALEXANDROWICZ et al 1981); sedimentation conditions altered on the flood-plains (MAMAKOWA 1970, RALSKA-JASIEWICZOWA 1980, ALEXANDROWICZ et al. 1981) and accumulation of tree trunks in alluvia (BECKER, SCHIRMER 1977). These stages are separated by quiet periods of relatively even discharge of the rivers, when mature meanders were formed.

The first humid period (8500-7500 BP) affected Podgrodzie in the Wisłoka valley in Poland (ALEXANDROWICZ et al. 1981) where the paleochannel fills were covered by the sands and silts after 8390$\pm$105 BP. In Tarnawa in the upper San valley the flood clay layer from between 8730$\pm$100 and 7840$\pm$100 BP was found in the peatbog (RALSKA-JASIEWICZOWA 1980). It was a period of landslides at Kamionka, the Carpathians in 8210$\pm$150 BP (GIL et al. 1974) and in loess areas at the bottom of ravines a paleosol was found dated at ca 8600 a BP (JERSAK 1977).

There is a Late Glacial paleochannel filled by silts and peats in the Vistula valley near Cracow at Nowa Huta at the foot of the scarp of Dłubnia alluvial fan (KALICKI 1984, 1987 and KALICKI, STARKEL 1987). In peat layers it is possible to distinguish a non-clayey lower level dated at 9660$\pm$ 1000 and the top from 8860$\pm$160 BP. The lower level is covered by clayey peats, often very compact. Clay in organic material proves that after about 8850 BP the river started to be more active.

On Rondo Mogilskie in Cracow a very similar sequence was found. The paleobotanic investigations by K. MAMAKOWA (1970) showed that littering silts come from the Younger Dryas and organic accumulation was interrupted here by sandy silts at the beginning of the Atlantic.

Several kilometres to the east, in the region of Zabierzów Bocheński on the site Drwinka, the accumulation of the Preboreal gyttjas and peats filling the paleochannel was interrupted by silts of high clay content. The top of the peats is dated at 7980±70 a BP (GĘBICA, STARKEL 1987).

Apart from the changes in sedimentation this period is marked by the recovery of channels. In Lasówka there is a system of three meanders filled by peats, and by clays on the top. The peaty fill bordering on sands in one of these meanders is from 7980±160 BP. It is probable that the avulsion of the Vistula channel occurred also during this period from the south of the contemporary Vistula to the northern side, at the foot of the scarp of the upper terrace where the Atlantic channels can be found.

The s e c o n d h u m i d p e r i o d occurred in 6500-5900 BP. It is known from the Wisłoka valley, where it is marked by the changes in channel system (ALEXANDROWICZ et al. 1981). A similar situation is observed in the Vistula valley in the vicinity of Cracow where there are numerous meanders from this period. In well investigated site at Pleszów (STARKEL et al. 1984, WASYLIKOWA et al. 1985) there is a paleochannel with several meanders at the foot of the scarp of the loess terrace filled by peat containing of deluvial horizons. The peat beds are dated 6255±40 BP. The paleochannel in Czyżyny has an analogous position somewhat to the West. It is filled by peat beds dated 6450±130 a BP. The paleochannel of the same period with fill formed 6700±130 a BP has been found in the Vistula valley in the Cracow Gate near Bielany (RUTKOWSKI 1987). The paleochannel deposits near the Wilga and Vistula confluence in the centre of Cracow are dated on the basis Fraxinus trunk at 6560±80 and 6340±120 years BP (SOKOŁOWSKI, WASYLIKOWA 1984). In this period the Vistula channel in the vicinity of Cracow and Nowa Huta underwent consecutive avulsions and shifted south from the loess terrace.

The n e x t h u m i d a n d c o l d p e r i o d came on the Atlantic/ Subboreal boundary, 5000-4500 BP. This period is marked in numerous sites by new peatbog formation (RALSKA-JASIEWICZOWA, STARKEL 1975). It was the period of rapid silt accumulation in valleys (FALKOWSKI 1975), of cutting paleomeanders (KOZARSKI, ROTNICKI 1977) or of resumed alluviation (FLOREK 1978).

In the Vistula valley near Cracow in the vicinity of Branice there is a paleochannel filled by silts (KALICKI, STARKEL 1987). At its base a trunk of Tilia sp. was found on the border with sands dated 5190±70 BP. Several kilometres to the east in the vicinity of the Grobla Forest there are abandoned paleochannel systems (STARKEL, KALICKI 1984; GĘBICA, STARKEL 1987). They are constituted of single meanders and less winding channel sections. In one of the older single meanders charcoal remnants in the fill were dated at 5420±110 BP (redeposition is possible) and a peat layer in the lower part of the fill dates back to 4860±110 BP. The middle part of the fill of the younger channel has an age of 4380±110 a BP. So, the Vistula river first extended its channel in the period of increased floods and then the avulsion of the Vistula channel to the north, up to the scarp of the loess terrace, took place.

The approximate dates from the paleochannels filled by silts were achieved in the vicinity of Jeziorzany in the Cracow Gate 4410±180 and 4010±180 a BP (RUTKOWSKI 1987) and also from the paleochannel deposits at the Wilga confluence where the Fraxinus trunk was dated at 4275±50 BP (SOKOŁOWSKI, WASYLIKOWA 1984).

This period is also marked in the earlier discussed site in Pleszów by accumulation of consecutive peat layers at the bottom of the flood-plain which was connected with rising ground water levels and dated 5380±60 to 4750±35 BP (WASYLIKOWA et al. 1985).

The last two periods of increased fluvial activity are conditioned not only by the climate but also by the human activity: mainly forest clearances and the development of agriculture.

The beginning of Subatlantic period in Poland is marked by growing humidity. It is connected with rising water levels in lakes e.g. innundation of the Łużyce culture village in Biskupin (SKARŻYŃSKA 1965), enlivened river activity and formation of new alluvial series dated 2700-2200 years BP at the Wisłoka river (ALEXANDROWICZ et al. 1981).

In the Vistula valley near Cracow this period is poorly recognized. However, in Cracow, in Aleja Pokoju, the paleochannel was found the peat fill of which is dated 2370±100 a BP. Also, in the Cracow Gate the paleochannel filled by silts was found and dated 2100±80 a BP (RUTKOWSKI 1987). It is characteristic that in the Branice-Stryjów gravel-pit a number of redeposited trunks were found one of which was dated 2220±120 a BP (STARKEL 1984; KALICKI, STARKEL 1987).

The period of increased activity of rivers in the 9th to 13th centuries is obviously connected with human activity. It was discovered that the channel facies in the Wisłoka valley from the 11th-16th centuries covers the older alluvia of flood-plain facies. This points to increased sediment transport from the deforested Carpathians (ALEXANDROWICZ et al. 1981).

The burial of cultural levels from the 10th-11th centuries under the silt layer, which marks the floods in the second half of the 11th century is known in Cracow in the Vistula valley (RADWAŃSKI 1972).

CONCLUSION

Having all the data in mind it is claimed that all climatic fluctuations during the Holocene are registered in the Vistula river valley in the vicinity of Cracow. In the period of the increased fluvial activity the consecutive new alluvial series were formed within the flood-plain (KALICKI, STARKEL 1987), there was a process of cutting-off single meanders and of avulsion of river channels. Organic sedimentation was interrupted by the accumulation of sandy of clayey silts, new peatbog formation on flood-plain, alluvial accumulation with tree trunks or burying cultural levels by flood-silts (Fig. 1).

The agreement between the curve of river regime changes drawn by L. STARKEL (1983b) for the rivers of Central Europe and the analogous curve drawn for the Vistula in the vicinity of Cracow is marked. The borderline dates in the Vistula valley for the two first periods are a little older: for the period on the Boreal/Atlantic boundary they are 8860±160 and 7980±160 a BP and for the Atlantic period 6700±130 and 6255±40 a BP. However, the period of increased activity on the Atlantic/Subboreal boundary seems a little longer since the border dates are between 5420±110 and 4010±180 a BP. The young periods in the Vistula valley are considerably poorer recognized and the anthropogenic changes play the important role in them. From the accumulated evidence Subboreal/Subatlantic boundary is limited by the dates 2370±100 and 2100±80 BP and the youngest period occurs at the end of 11th century.

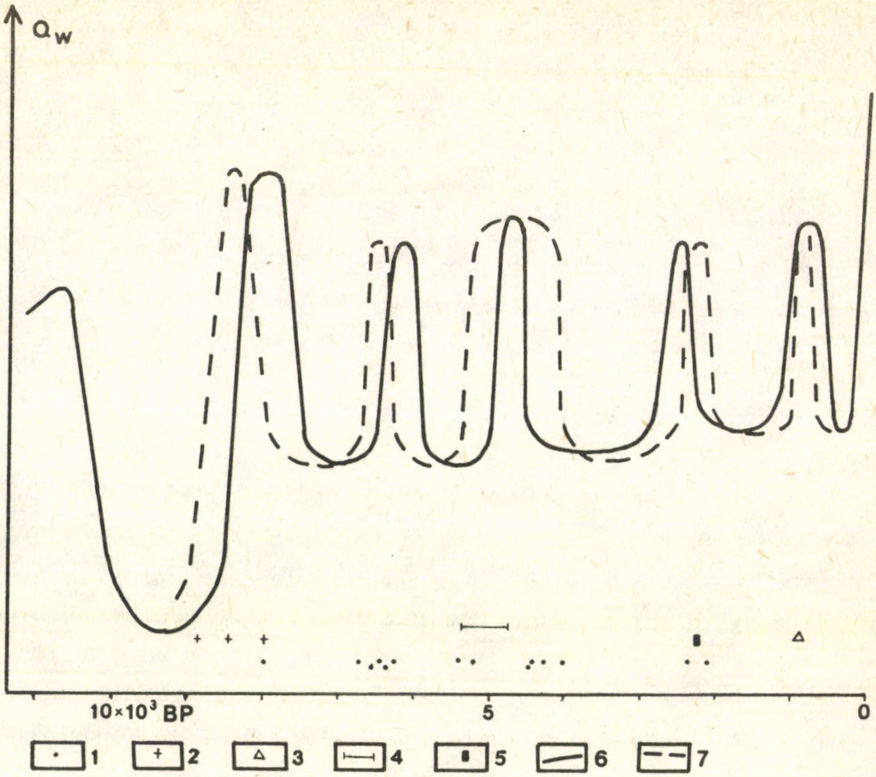

Fig. 1: Changes in river discharge and a list of C^{14} dates from the Vistula valley in the vicinity of Cracow.
1 = cut-off paleochannels and their system 2 = break in organic sedimentation; 3 = burial of cultural levels by silts; 4 = periods of peat accumulation; 5 = tree trunks; 6 = changes of river discharge in Central Europe during the Holocene (STARKEL 1983b, 1986); 7 = changes of discharge of the Vistula in the vicinity of Cracow during the Holocene.

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DISTINCTION OF VISTULIAN AND HOLOCENE FLUVIAL DEPOSTIS

E. Niedziałkowska

ABSTRACT

The grain size and grain roundness of the Vistulian and Holocene alluvia on the margins of the flysch Carpathians are treated in the paper. Deposits 3 selected sites in the Wisłoka valley and 3 sites in the upper Vistula region /Fig. 1/ are analyzed. The study areas differ with respect to the catchment size, geological structure and hydrological regime. Various coefficients reflecting sedimentation environments are presented.

Fig. 1: Location of study area

INTRODUCTION

Facial and stratigraphic aspects of sediments in the Wisłoka valley /Fig. 1/ have been published in the monograph edited by L. STARKEL /1981/. Similar characteristics of fluvial deposits in the upper Vistula valley /Fig. 1/ were investigated by author /GILOT et al. 1982; NIEDZIAŁKOWSKA et al. 1985/. The aim of this paper is to complete the latter characteristics with the parameters of grain size and roundness.

Grain size parameters as mean diameter /M_Z/, standard deviation /σ_I/, skewness /S_{KI}/ and kurtosis /K_G/ were calculated according to the formulae of FOLK and WARD /1957/. An analysis of quartz grain roundness /grain diameter 0.75-0.5 mm/ was performed using an automatic graniformometer. Two indices have been calculated: 1. roudness coefficient /W_o/ according to KRYGOWSKI's formula /1964/ and 2. standard deviation /σ_o/ as an indicator of uniformity.

The sediments from various terrace levels have been analysed and dated by the C^{14} method. The sediment facies have been classified based on the interpretation of the outcrops, sedimentation structures, and grain size.

The term "facies" understood as a type of sediment of given properties has been used by the author to determine the environment of the sediment deposition. The following facies have been distinguished after ALLEN /1965/

a. channel facies: sediments deposited in the channel;
b. flood-basin facies: material deposited on the floodplain outside the levee;
c. channel-fill facies: mineral or organic material deposited in abandoned channels;
d. alluvial fan facies: accumulation of sediments deposited by a tributary.

GRAIN SIZE

Channel facies PL1, PL2, LG-AL, SB-SA* of the Wisłoka valley presented in the graphs of the relation between grain size coefficients are charac-

*The conventional abbreviations of time units for the Vistulian and Holocene

PL1 - Older Pleniglacial
IPL - Interpleniglacial
PL2 - Younger Pleniglacial
LG - Late Glacial
AL - Alleröd
PB - Preboreal
BO - Boreal
AT - Atlantic
SB - Subboreal
SA - Subatlantic
SA2 - the last ca 1000 years

terized by similar values of mean diameter, sorting, skewness and kurtosis /Fig. 2/. They have poorly sorted gravel - and - sand fractions as well as sand fractions of good sorting. The positive skewness indicates evironments of decreasing flow velocities, while the K_G value indicates stable conditions of deposition.

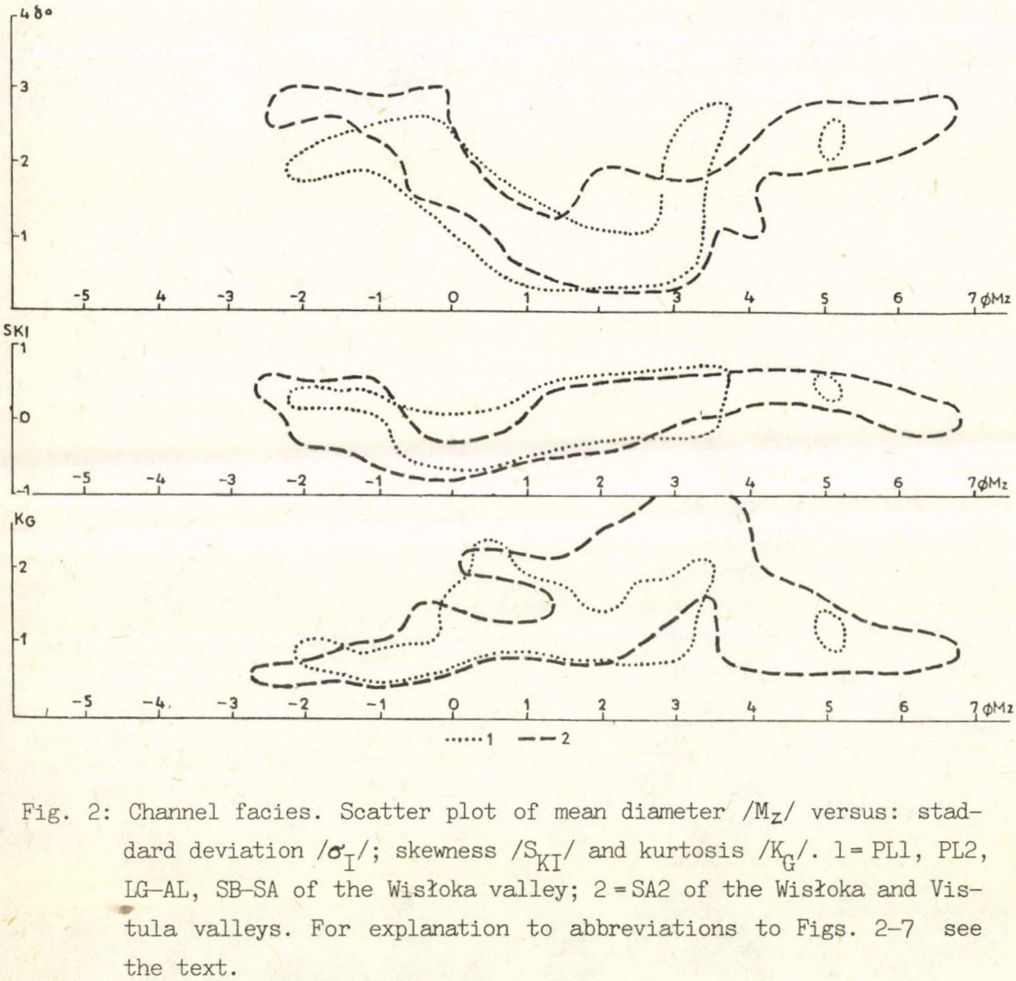

Fig. 2: Channel facies. Scatter plot of mean diameter /M_Z/ versus: staddard deviation /σ_I/; skewness /S_{KI}/ and kurtosis /K_G/. 1 = PL1, PL2, LG–AL, SB–SA of the Wisłoka valley; 2 = SA2 of the Wisłoka and Vistula valleys. For explanation to abbreviations to Figs. 2–7 see the text.

The channel facies SA2 of the Wisłoka and Vistula valleys /Fig. 2/ have, besides the pattern described above, a larger amount of sediments with smaller mean diameters and very poor sorting. These are silty-sandy layers. The positive skewness as well as K_G values provide that they were deposited in non-current environment.

The author interprets the definite differences between the former and latter facies as a proof of variable hydrodynamic conditions in the channel. The older channel facies /Vistulian, LG-AL/ have been deposited by rivers of slightly different discharges. While the sedimentation of the younger channel facies /SA2/ was accompanied by strongly different discharges.

Flood-basin facies PL2, AT, SA2 of the Wisłoka and Vistula valleys are characterized by small mean diameters and very poor sorting /Fig. 3/. The positive skewness and K_G values provide evidence of non-current environment.

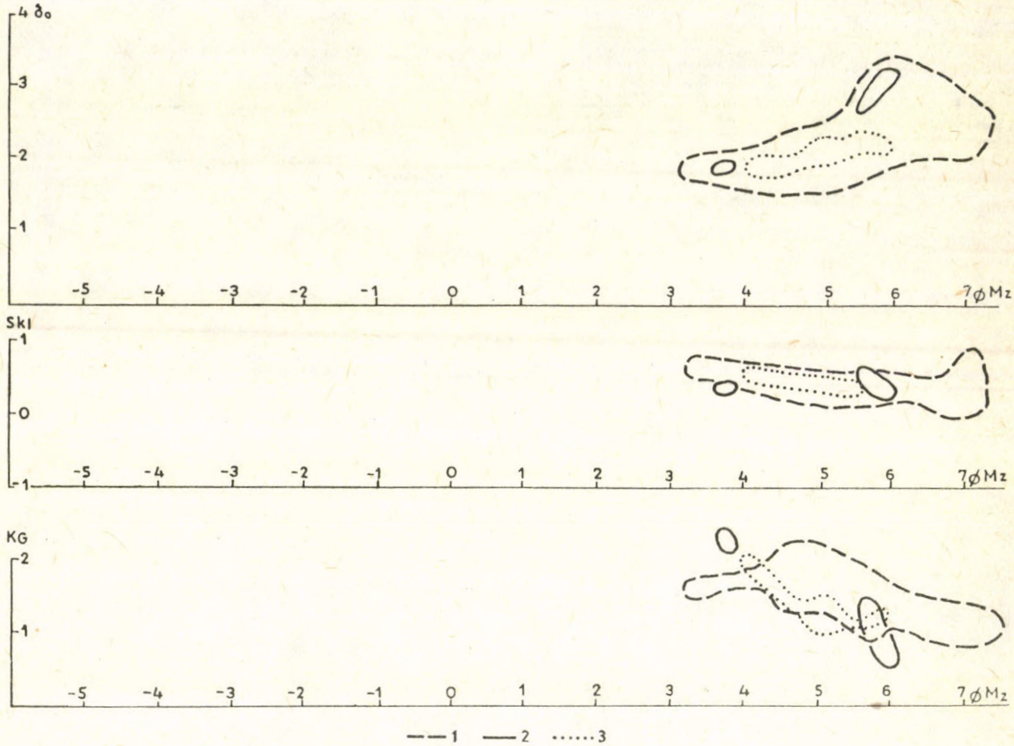

Fig. 3: Flood-basin facies. Scatter plot of mean diameter /M_z/ versus: standard deviation /σ_I/; skewness /S_{KI}/ and kurtosis /K_G/. 1 = PL2 of the Vistula valley; 2 = AT of the Wisłoka valley; 3 = SA2 of the Vistula valley

Channel-fill facies IPL, PL2, LG-AL, PB-BO, SA2 of the Wisłoka valley and those LG-AL of the Vistula valley do not show different grain size coefficients with respect to their age /Fig. 4/. However, significant variation in mean diameter and in sorting reflect various sedimentation conditions existing in abandoned channels. The coarse-grain deposits possess the properties of channel facies. This indicates periodic activity in the channel. The fine-grain sediments exhibit properties of flood-basin facies. They were deposited by precipitation of suspended matter in non-current environment.

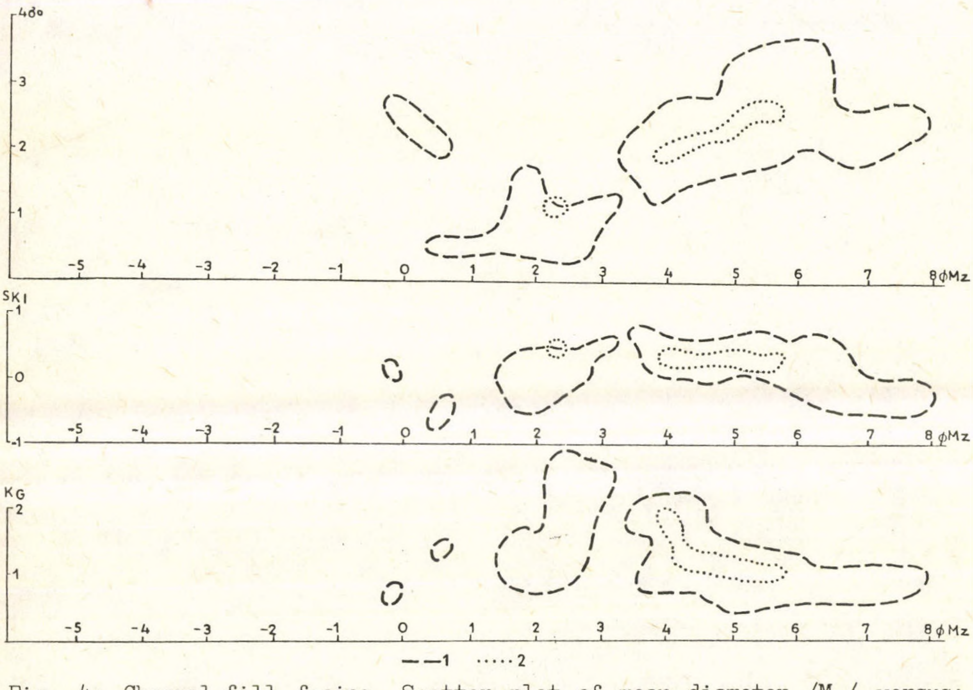

Fig. 4: Channel-fill facies. Scatter plot of mean diameter /M_z/ versus: standard deviation /σ_I/; skewness /S_{KI}/ and kurtosis /K_G/. 1 = IPLo, PL2m, LG–ALm, PB–BOo, SA2m of the Wisłoka valley; 2 = LG–ALm of the Vistula valley

The alluvial fan facies AT, of the Wisłoka valley, like channel-fill consists of two various groups of deposits /Fig. 5/. The first sandy group with a small variation in mean diameter and moderate sorting, was deposited on an alluvial fan surface within the channel range. The second group, with finer mean diameter, and poor to very poor sorting, is characterized by

K_G values above 1. The latter indicate the occurrence of this deposit in environments of higher dynamics. The positive skewness suggests a propensity to velocity decrease. Sediments with such properties could have been deposited in a zone located close to the channel.

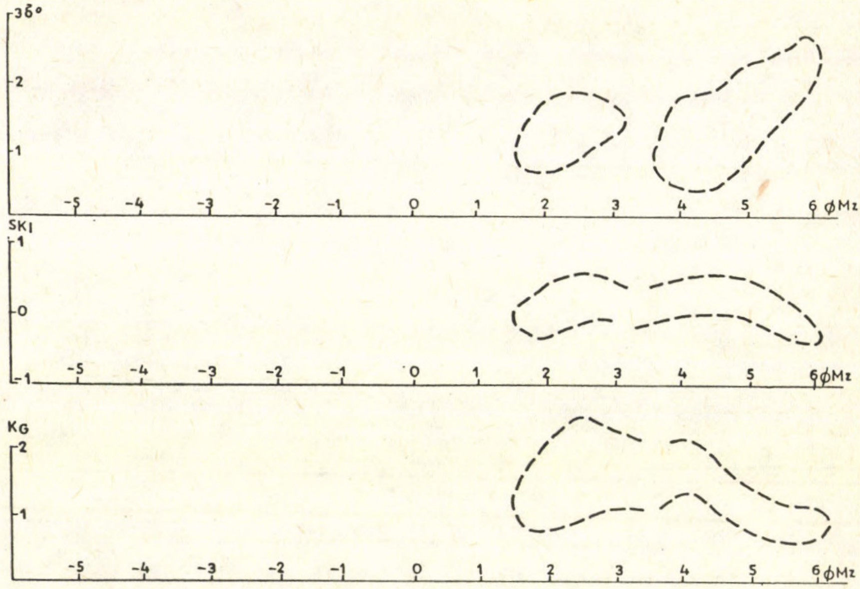

Fig. 5: Alluvial fan facies. Scatter plot of mean diameter $/M_z/$ versus standard deviation $/\sigma_I/$; skewness $/S_{KI}/$ and kurtosis $/K_G/$ of the Wisłoka valley of AT period

GRAIN ROUNDNESS

Despite of their different ages, the deposits of channel facies of the Wisłoka and Vistula valleys are characterized by similar values of the W_o coefficient /Fig. 6a/. However, the uniformity coefficient varies. In the case of the older channel facies, it is generally less variable than in the case of the younger ones.

The flood-basin facies both of the Wisłoka and Vistula valleys are characterized by low values of the W_o coefficient /Fig. 6b/. It is worth to note that the deposits of the flood-basin facies of AT are characterized by the larger values of W_o than in the case of those deposited in SA2. Author seeks the explanation of this phenomenon in variable sedimentation conditions on the flood-plains in these periods, in a distance from the channel, and in a high frequency of floods.

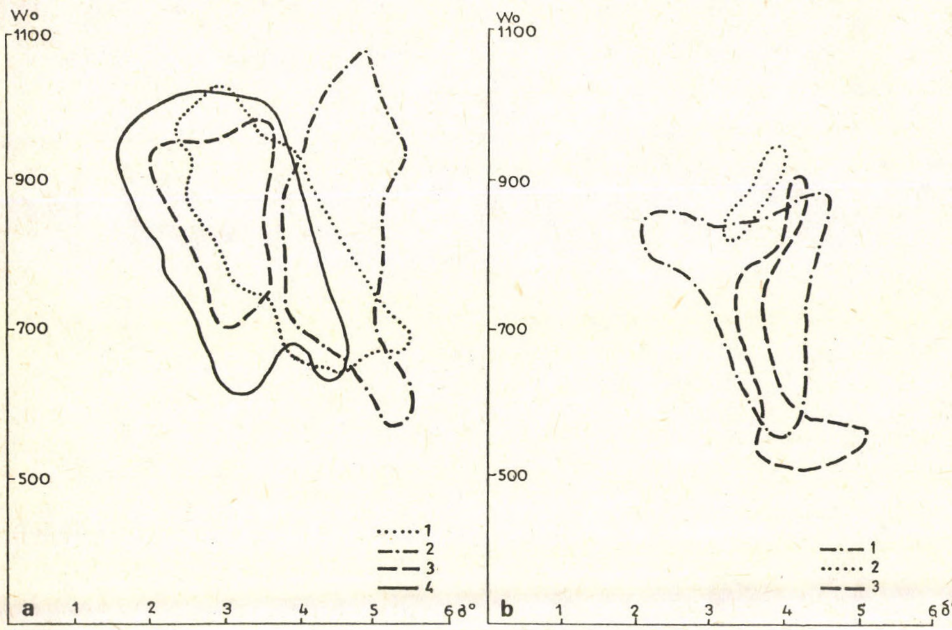

Fig. 6: Scatter plots of roundness coefficient /W_o/ versus uniformity index /σ_o/. a = Channel facies. 1 = PL1, LG-AL; 2 = PL2; 3 = SB-SA of the Wisłoka valley; 4 = SA2 of the Wisłoka and Vistula valleys. b = Flood-basin facies. 1 = PL2 of the Vistula valley; 2 = AT of the Wisłoka valley; 3 = SA2 of the Vistula valley

The channel-fill facies of the Wisłoka and Vistula valleys, both the mineral and the organic sub-facies belonging to different periods of the Vistulian and Holocene, form in the graph a compact cluster and their W_o coefficients are poorly differentiated /Fig. 7a/. Only the mineral sub-facies of SA2 is slightly better rounded and more uniform. These are the properties of young Holocene channel facies.

Alluvial fan facies characterized by better-rounded grains than channel facies deposits /Fig. 7b/. The range of the uniformity coefficient values is also larger than in the case of other facies. It is shifted towards lower uniformity, in general.

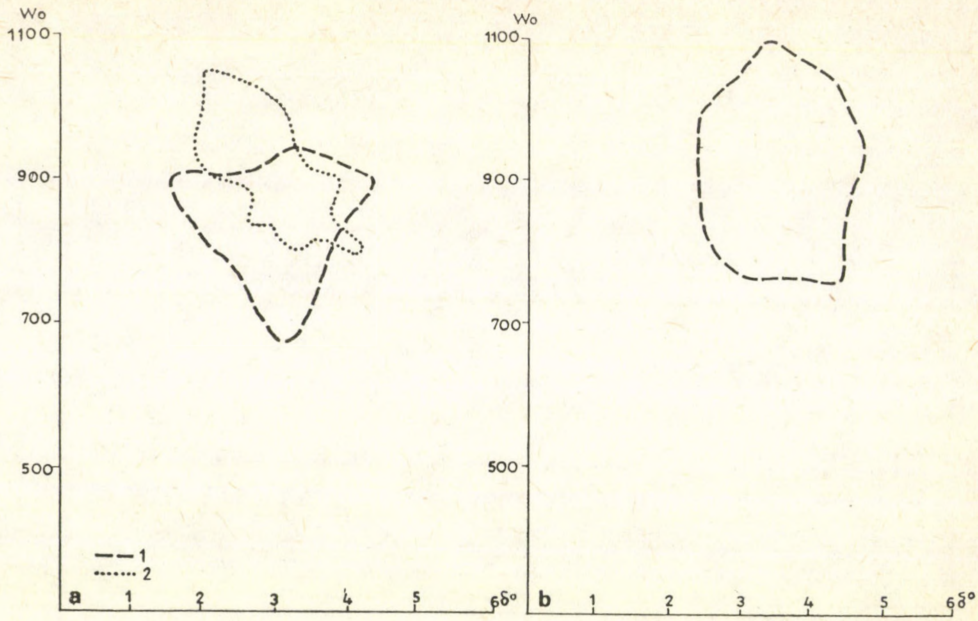

Fig. 7: Scatter plots of roundness coefficient /W_o/ versus uniformity index /σ_o/

a = Channel-fill facies. 1 = IPL, PL2, LG–AL, PB–BO of the Wisłoka and Vistula valleys; 2 = SA2 of the Wisłoka valley. b = Alluvial fan facies of the Wisłoka valley of the AT period

CONCLUSIONS

1. The values of grain size and roundness coefficients are not subjected to the changes in particular periods of the Vistulian and Holocene, excluding those referring to the sediments deposited in SA2, the latter in author's opinion reflect human impact.
2. The values of grain size and roundness coefficients are not controlled by geological structure.
3. The grain size coefficients reflect the conditions of sedimentation environment.
4. The differences between the W_o coefficient values in the selected facies in author's opinion, result from grain morphoselection only, in agreement with KANIECKI /1976/ and MŁYNARCZYK /1985/ findings.

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